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WORKSHOP ON

ANCIENT CRUSTS OF THE TERRESTRIAL PLANETS

Houston, Texas
12-14 February 1979

A LUNAR AND PLANETARY INSTITUTE WORKSHOP

UNIVERSITIES SPACE RESEARCH ASSOCIATION
LUNAR AND PLANETARY INSTITUTE
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WORKSHOP ON
ANCIENT CRUSTS OF THE TERRESTRIAL PLANETS

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A LUNAR AND PLANETARY INSTITUTE WORKSHOP

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INTRODUCTION

This volume contains abstracts of talks presented at a Workshop on Ancient Crusts of the Terrestrial Planets held at the Lunar and Planetary Institute on 12-14 February 1979. Related materials, also found in this volume, include a summary of the workshop, program, participant list, and a short bibliography.

The conveners of this workshop were Dr. James J. Papike (State University of New York, Stony Brook), Dr. Charles H. Simonds (Northrup Services, Inc., Houston), and Dr. Thomas R. McGetchin (Lunar and Planetary Institute).

Logistical and administrative support was provided by Carolyn Kohring (Research/Administrative Assistant, LPI), who also compiled this volume.

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SUMMARY OF WORKSHOP ON ANCIENT CRUSTS OF THE TERRESTRIAL PLANETS

T. R. McGetchin, C. H. Simonds, P. W. Weiblen and J. Wooden

DESCRIPTION OF THE WORKSHOP AND BACKGROUND

On 12-14 February 1979, approximately 50 scientists participated in a workshop on "Ancient Crusts of the Terrestrial Planets," held at the Lunar and Planetary Institute. The workshop encouraged interaction between scientists studying the ancient terrestrial rocks and those studying the old crusts of the planets of the inner solar system.

The Earth

The workshop opened with reviews of the present data base on the geology and geochemistry of the Archean as exposed in North America (the Superior Province, Labrador, and Wyoming) and Africa (the Kaapvaal craton).

The Superior Province represents the largest and best exposed example of an Archean greenstone-granite terrane; however, the discussion emphasized that models of its evolution remain conjectural. The principal uncertainities appear to be the following: 1) What was the tectonic environment in which the greenstone-granite terrane formed (rift, back arc basin, plate subduction, sag subduction)? 2) Did the greenstone-granite terrane develop on older "continental" crust? 3) If an older "continental" crust was present, what was its composition, age, extent, and involvement in magma production? 4) What were the original relationships between the greenstone-granite terranes and the gneiss terranes? 5) Under what thermal gradients and convection conditions did greenstone-granite terranes develop? 6) What were the original thicknesses of the greenstone volcanic sequences? 7) What is the ultimate origin and nature of the 2.8-2.5 b.y. event that is common to many of the Archean cratons?

The combination of detailed mapping and isotopic studies in the Saglek-Hebron area in Labrador provides a well-constrained interpretation of the origin of a complex, high-grade Archean gneiss terrane. A succession of crustal additions and thermal and tectonic events have been deciphered which extend from 3.8 to about 1.7 b.y.

It was suggested on the basis of trace element data that the tonalitic rocks of the greenstone-granite terrane of northern Minnesota are derived from a basaltic source which in turn was derived by partial melting of the mantle. The tonalitic to adamellitic rocks of the Minnesota River Valley do not have a basaltic source, but instead require a light REE enriched source whose bulk composition may be roughly dioritic. It was emphasized that a crustal source region such as that suggested for the Minnesota River Valley rocks is so large that it probably contains a number of rock types whose <u>average composition</u> must be considered in partial melting processes.

A very similar but not identical sequence of events is documented for southwestern Greenland. Rocks, greater than 3.0 b.y. old, are also present in northern Michigan, Wisconsin, and southern Minnesota. This dominantly gneissic terrane, which lies south of the southernmost exposures of the Superior Province greenstone-granite terrane, also records a complex history ranging from 3.6-1.7 b.y. with major events of roughly the same age as those of Greenland and Labrador. The Laramide uplifts of Wyoming and Montana contain several Archean sequences which current research indicates are dominated by 3.0 b.y. and younger Archean rocks. The Archean rocks of the Kaapvall craton, southern Africa were described. It was noted that the major greenstone-granite terrane of the Kaapvaal is older than that of Canada and western Australia.

The isotopic record of Archean rocks is complicated by the complex thermal and tectonic events that have occurred during the stabilization of the older crust. The Rb-Sr radiometric system is the most widely applied isotopic system in Archean rocks, but it is subject to many problems including Rb loss during recrystalization, Rb gain in relative low Rb systems by diffusion or fluid transfer during metamorphism, Rb gain during periods of intrusion of high alkali magmas and/or associated fluids, and Sr loss during shearing. Previously determined initial Sr ratios for the older Archean rocks of Labrador and Greenland may be slightly high as new studies of the oldest identified Labrador rocks indicate ages of about 3.85 b.y. and Sr initials of 0.7000-0.7005. Sr isotopes do indicate that the majority of the Archean crust was developed in short (50-200 m.y.) episodic cycles that involved the multi-stage processing of material (probably basalts) derived from the mantle. The combined use of the Rb-Sr, Sm-Nd, Pb-Pb, and U-Pb zircon systems should provide a powerful tool for determining Archean chronologies and in addition provide much data useful for constraining petrogenetic processes during the Archean.

Thickening of the crust in the Archean occurred in response to a unique but poorly understood thermal regime. The double convective systems of Mackenzie and Weiss were discussed. A heat flow model with a maximum of about twice the present value at 2.8 b.y. and a subsidiary maximum at 3.5 b.y. was suggested and related to extensive production of scattered sialic crust in this interval.

It is apparent from the structural record that stabilization of the crust in Archean greenstone-granite terranes involved a complex tectonic regime. Modern-day plate tectonic analogs can be found for many of the tectonic features of greenstone-granite terranes; however, detailed field studies in the western part of the Superior Province reveal two, possibly overlapping, periods of tectonic deformation: an early period responsible for the gross structure and folds within the metavolcanic terrane and a second period related to large-scale transcurrent faults. The former can be related to gravity tectonics associated with emplacement of ($\sim$50 km) tonalitic diapirs and the latter with horizontal motion tectonics.

Understanding the tectonics of the Archean represents a major challenge. The discussion emphasized the need for caution in using a simple uniformitarian approach to the use of modern day analogs to a more mobile and thinner lithosphere. The problem is further complicated by the fact that impact cratering may have played a significant role in establishing fracture patterns and localizing igneous activity and sedimentation. In view of the record of meteorite impact on the other inner planets, this process must be taken into account in the early (4.5-3.9 b.y.) thermal evolution of the earth.

The Moon

Among the terrestrial planets the moon represents the simplest example of an evolved body. The essence of petrologic, geochemical and isotopic studies of the moon comes from the fact that the planet virtually lacks any H_2O or CO_2 or any evidence for tectonic processes other than those driven by impact. Chemical fractionation on the moon is assumed to take place almost exclusively by igneous processes (crystal settling and partial melting). The lack of water or an atmosphere precludes fractionation by weathering and sedimentation.

Also the lack of water probably precludes the kinds of chemical migration often inferred to occur in terrestrial metamorphic terranes. Although impact induced fractionation by volatilization has been suggested numerous times, it has not been unequivocally demonstrated to operate in any environment other than that in the upper few meters of lunar soil, and is not relevant to the kilometer and larger scale impacts which dominate the moon's cratering record.

Study of the Apollo samples established that the outer few hundred kilometers of the moon were fractionated by igneous processes during the first few hundred million years following accretion. Direct evidence for this event includes a small number of coarse grained anorthosites, troctolites, norites and dunites, which in some cases have cumulate textures and which in two cases have yielded crystallization ages of 4.2-4.6 b.y. Many of the other non-mare samples yield isotopic data indicating fractionation of Rb from Sr very early in the planet's history. Further evidence for the early fractionation comes from trace element studies of the mare basalts (crystallization ages 3.9-3.0 b.y.) which suggest that they come from a source which itself fractionated from feldspar. Modeling of Sm-Nd, Rb-Sr and U-Pb isotopic systematics of the mare basalts suggests that the early fractionation of the source of the mare basalts also took place early in lunar history. The process of the early fractionation has been conceptualized by some workers as occurring by crystal fractionation of a planet encircling ocean or magma which has an ultramafic (almost but not quite chondritic) source. Much of the controversies in describing early fractionation involve the details of isotopically dating the samples, modeling the isotopic data, and understanding the best ways to interpret trace element analyses of coarse grained plutonic rocks and arrive at a reasonable estimate of the composition of the primary liquid from which they crystallized.

Impact processes have extensively reworked the lunar surface, and have introduced several percent meteoritic contamination, as indicated by the level of siderophyle contamination in impact produced breccias above endogenous lunar level. The precise dating of the impact events represented in the Apollo collection, and by inference the age of the bulk of the morphologic craters visible in the highlands is difficult analytically, but of great importance, since these ages form the basis of dating impact events throughout the inner solar system.

Volatiles — A Major Variable

It is concluded by most workers that one of the principal controls on the igneous fractionation of the planets is the H_2O and CO_2 contents. Partial melting of ultramafic compositions is strongly affected by the amount of water and CO_2 in the source and phases containing those volatile species. Adding water to dry peridotite generally increases the degree of silica saturation and any partial melts, as well as producing a major reduction in the temperature required to initiate melting. Carbon dioxide also has strong effects on altering the melting temperature and composition of basalts derived. Thus understanding the amount and ratio between the various potential volatile species is critical to understanding planetary fractionation.

CONCLUSIONS

The Questions

What is the relationship between the ancient lunar uplands and the Archean of the earth? Were the ancient crusts of all the terrestrial planets floated out of a "magma ocean" as some lunar scientists postulate? Or is water so important in terrestrial petrogenesis that the moon and earth have little or nothing to do with each other, as far as their crusts are concerned.

In jumping into such strange waters, we found many fundamental questions had to be asked — few had answers, but it was good to begin defining the issues. Among the most prominent questions are: (1) what is the nature of the oldest terrestrial silicic crust accessible for study — its composition, structure and origin? What are the precursor (source) rocks from which the earliest silicic rocks were derived? (2) What is the origin of the oldest volcanic rocks; are they related to plate margin processes, mid-plate features (similar to Hawaii) and do they represent a tectonic style (thin lithosphere) unique to the Archean? (3) What is responsible for the episodes of petrogenesis in the Archean? Does this episodicity reflect internal convection? Are the episodes real? (4) What is the nature of the lower crust of the earth? Is water pervasive there? How did the lower crust originate? How is it related to the upper crust; the upper mantle? Is it currently being formed or does it represent an early terrestrial crust which developed directly on top of the Moho? (5) How did the lunar uplands originate? Did a similar process work on the earth? Would a process of plagioclase floation work in volatile-bearing magma? (6) How important were (and are) volatiles in the formation of the earth's crust? What are the volatiles in the lower crust? Where did they come from? (7) How is water distributed throughout the solar system — that is, are planets nearer the sun likely to contain less water (and other volatiles) because they accreted in a hotter environment? (8) What about the other planets? What do Mars, Venus and Mercury tell us about the earth's crust? What do we really know about these surfaces? Among the array of terrestrial planets, the earth is the largest, the moon the smallest — so if processes differ because of planetary mass, then one might think so. There may be other more important ones, the abundance of water for example.

Present State of the Answers

The Moon

The Moon has a crust which varies in thickness from 2 to 60 km and is older than 4 b.y., in general. Upland samples returned by Apollo 14, 15, 16 and 17 suggest that it is anorthosite and norite, pervasively brecciated. It is commonly believed these feldspar-rich rocks floated to the surface of a primordial circumlunar magma-ocean. Lunar crustal petrogenesis was *anhydrous*.

The Earth

The earliest crust of the Earth is nowhere preserved. The oldest rocks are granodioritic or tonalite gneisses, believed to be derived by partial melting of a yet older tonalitic (?) parent, perhaps in the lower crust.

The Other Planets

Data are very sparce, but . . .

Future Work

Resolution of the issues raised will require much more data than we presently have at our disposal. For the earth, we need volume estimates of the major rock units, the sequence of events responsible for their formation and their stress and structural history. This will require field observation and extensive laboratory work on carefully selected suites of rocks. Data from the other planets will be provided by the planetary exploration program, but little of much significance for this problem can be anticipated for more than a decade.

Valuable new data utilizing existing techniques can be expected from (1) observational field petrology, (2) experimental petrology, (3) process analog models and (4) theoretical modeling. Possible *new* approaches and techniques which may provide important new insights are: (1) coordinated studies on well selected suites of samples (such as has been accomplished on the lunar samples), (2) geophysical consortia (such as COCORP) which may provide needed data on regional structural relationships, (3) continental drilling, (4) possibly observations of the earth from space, and (5) comparative planetology.

Workshop on
ANCIENT CRUSTS OF THE TERRESTRIAL PLANETS

PROGRAM

Monday, 12 February 1979

Welcome
Tom McGetchin

Objectives of Workshop
Chuck Simonds

REGIONAL GEOLOGY, GEOCHEMISTRY AND GEOPHYSICS OF THE TERRESTRIAL ARCHEAN

Archean of Manitoba and Comparisons with Other Archean Areas
Lorne Ayres

Archean of the Southern Part of the Canadian Shield
Paul Sims

Early Archean Crust of Northern Labrador
Ken Collerson

Geochemistry of the Southern Part of the Canadian Shield and the Origin of Tonolitic Magmas
Gil Hanson

Geochemistry of the Kaapvaal Craton
Fred Barker

Archean Age Relationships
Sam Goldich

Isotopic Constraints on the Evolution of the Archean Crust
Joe Wooden

Interpretation of Scattered Isotopic Data from the North Atlantic Craton
Ken Collerson

GEOLOGICAL AND GEOPHYSICAL INSIGHTS INTO THE NATURE OF THE LOWER CRUST

Xenolith Studies and the Nature of the Lower Crust
Bob Kay

Seismic Structure of the Lower Crust
Larry Brown

Seismic and Other Geophysical Information about the Lower Crust and Upper Mantle
Don Hall

Electromagnetic Sounding of the Lunar and Terrestrial Crust
Dave Strangway

Tuesday, 13 February 1979

LUNAR HIGHLANDS GEOLOGY AND RELATED TOPICS

A Synthesis of Lunar Highlands Petrology, Geochemistry and Geochronology
Chuck Simonds

Origin of Lunar Anorthosites
John Longhi

Layered Igneous Intrusions and Anorthosites and the Processes of Fractionation
Stewart McCallum

Igneous Fractionation of the Lunar Highlands Crust
Gordon McKay

The Marcy Anorthosites: A Comparison with Lunar Anorthosites
Larry Haskin

Phase Equilibria of Silicates as a Function of P(O_2, H_2O, CO_2 and other Volatile Species)
Art Boettcher

TECTONICS OF THE TERRESTRIAL PLANETS

Observations of the Tectonics of the Terrestrial Planets
Jim Head

Archean Shield Structure at the Surface and the Role of Gravity Tectonics
W. M. Schwerdtner

Plate Tectonics on the Early Earth
Bill Kidd

What Impact Can and Cannot Do to Planetary Crusts
Dick Grieve

History of Heat Flow and Convection in the Earth
Richard Lambert

Review of Tectonic Processes in the Inner Solar System
Paul Lowman

The Role of Water in Planetary Tectonics
Jeff Warner

Wednesday, 14 February 1979

An informal discussion of possible mechanisms to further cooperative research between scientists studying the terrestrial archean and those studying the moon and planets.

Workshop on
ANCIENT CRUSTS OF THE TERRESTRIAL PLANETS

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CRUSTAL DEVELOPMENT OF THE WESTERN PART OF THE ARCHEAN SUPERIOR PROVINCE, MANITOBA AND NORTHWESTERN ONTARIO. L.D. Ayres, Department of Earth Sciences, University of Manitoba, Winnipeg, Manitoba, Canada R3T 2N2

The dominant feature of the western Superior Province is a series of east-trending subprovinces that differ in lithology, structure, and metamorphic grade. The subprovinces range in width from 40 to 300 km and represent deformed primary lithologic sequences that are in part now fault-bounded. Seven subprovinces have been defined. From south to north these are: Wawa metavolcanic-granodiorite (V.G.), Quetico metasedimentary, Wabigoon V.G., English River metasedimentary-granitoid, Uchi V.G., Berens River granitoid, and Sachigo V.G. subprovinces. The variations among these subprovinces probably reflect different processes of crustal evolution.

The four metavolcanic-granodiorite subprovinces comprise isoclinally folded metavolcanic-metasedimentary sequences (greenstone belts) intruded by large, composite, diapiric granodiorite-trondhjemite batholiths. Metamorphic grade in the greenstone belts ranges from lower greenschist in the centre to lower amphibolite at the margins. Each greenstone belt is only a remnant of the original volcanic-sedimentary sequence which was dismembered by emplacement of the batholiths.

These subprovinces appear to represent deformed linear volcanic island chains that developed along major fractures or rift systems in the early crust, about 2.75 to 2.7 Ga. Initial volcanism produced a series of coallescing, subaqueous, komatiitic and tholeiitic basaltic shield volcanoes that gradually built up toward sea level. Extensive downsinking of the volcanoes due to isostatic loading of the crust resulted in eruption of thick ($>$ 10 km) subaqueous pillowed basalt sequences in oceans only 2 to 3 km deep. The shield stage was superceded by large subaerial and subaqueous, largely pyroclastic, calc-alkaline dacitic to rhyolitic stratovolcanoes. Much of the subaerially erupted tephra was reworked and eventually deposited on the subaqueous flanks of the volcanoes as volcaniclastic greywacke and conglomerate. When the reworked component is considered, Archean volcanism produced a distinctly bimodal basalt-dacite suite. This bimodality implies a dual magma source; basalt from the mantle and dacite from partial melting of the sinking basaltic base (now amphibolite) of the volcanoes.

The extensive batholiths are 15 to 20 km thick sheets that may represent a deeper-seated continuation of the magmatism that produced the stratovolcanoes. This major plutonic event stabilized the present crust and further developed the tight isoclinal folding that had been initiated during downsinking.

The pre-volcanism crust was largely destroyed by emplacement of the younger batholiths, but the identified remnants are largely trondhjemite-tonalite plutons (2.95-3.05 Ga), metamorphosed to amphibolite facies. The nature of the supracrustal sequence into which the plutons were emplaced is unknown, but the oldest known volcanic sequence (2.95 Ga) is about the same age as the plutons and may have been part of the early crust. The presence of the trondhjemite may imply an earlier mafic (volcanic?) event which would have been the partial melting source of the trondhjemite magma. Prior to the extensive 2.75 Ga volcanism, the early Archean crust must have been uplifted and eroded to expose the plutons, and consequently formed continental land masses of unknown extent. When the 2.75 Ga volcanism

CRUSTAL DEVELOPMENT OF THE WESTERN SUPERIOR PROVINCE

Ayres, L.D.

began, much of the early crust was submerged, although local crust-derived sedimentary units attest to some emergent crust.

The Quetico and English River subprovinces differ from the metavolcanic-granodiorite (V.G.) subprovinces in the rarity of metavolcanic units, abundance of meta-turbidite, higher metamorphic grade (amphibolite to locally granulite facies) and lower abundance of granitoid rocks in most areas. The meta-turbidites appear to be a facies equivalent of the dacitic volcanism in the adjacent subprovinces and represent deposition in the open ocean or in linear basins between volcanic island chains. Reconstructed facies relationships across the Quetico subprovince indicate that the volcanic island chains developed sequentially rather than simultaneously.

The higher metamorphic grade of the Quetico and English River subprovinces compared to the V.G. subprovinces is anomalous. One would expect that the concentration of magmatism in the V.G. subprovinces would have resulted in a higher metamorphic grade than in the adjacent sedimentary basinal deposits. The low metamorphic grade of the V.G. subprovinces may reflect suppression of metamorphic isograds by rapid downsinking of the volcanoes.

The Berens River subprovince represents a deeper, more plutonic crustal level, but it is poorly known.

There are many similarities in volcanism, sedimentation, plutonism and metamorphism between the Superior Province and other Archean shield areas. In all shields extensive greenstone belt volcanism followed by plutonism was a major crust stabilization event. However, it occurred at different times in different shields.

TWO ARCHEAN GRAY GNEISS COMPLEXES: KAAPVAAL CRATON AND BIG HORN MOUNTAINS, Fred Barker, U. S. Geological Survey, Denver, CO 80225

Archean gray gneiss complexes, except for their included remnants of supracrustal rocks, form the oldest known terrestrial crust. In many localities they are found adjacent to or may occur as basement to younger greenstone belts of the classic younger Archean type. These gneiss terranes consist largely of complexly deformed, banded tonalitic and trondhjemitic gneisses (IUGS classification), of minor to abundant metabasalt of tholeiitic, calc-alkaline and not uncommon komatiitic types, of minor post-tonalitic-trondhjemitic-gneiss intrusive quartz diorite to granodiorite, and rare metasedimentary rocks. A comparative study of 3.0-g.y.-old gray gneiss complexes of the Big Horn Uplift, Wyoming, and of the pre-3.4-g.y.-old Ancient Gneiss Complex of the eastern Kaapvaal Craton, Swaziland and Transvaal, shows many similarities as well greater complexities in the more extensive African terrane.

The Wyoming gneisses are well banded in biotite versus plagioclase and quartz, at least twice deformed, and metamorphosed to upper amphibolite facies. They are trondhjemitic, of high-Al_2O_3 type (>15%), HREE-depleted, show Rb/Sr ratios of about 0.1 and an initial $^{87}Sr/^{86}Sr$ ratio of 0.7000. Small (<20 m wide) lenticular inclusions of metabasalt form 1 - 2% of the complex: of five samples analyzed to date, one is primitive and LREE depleted, three are of typical olivine-normative Archean tholeiite, and one is quartz-normative, moderately LREE-enriched tholeiite. The Ancient Gneiss Complex consists of an older bimodal suite, a younger metamorphite suite, and siliceous intrusives of several types. The bimodal suite includes closely interlayered, largely trondhjemitic gneisses of garden variety high-Al_2O_3, HREE-depleted type and of the rather rare low-Al_2O_3, high-SiO_2 (>75%), LREE-enriched and non-HREE-depleted type, as well as metabasalt ranging from komatiitic to calc-alkaline types. The metamorphite suite consists largely of microcline gneiss, metatholeiite and low-K siliceous gneiss.

The preferred mode of generation of these ancient gray gneiss complexes, following the suggestion of T. H. Green and Ringwood (1968) and as presented by Barker and Arth (1976), involved accumulation of thick piles of basalt, metamorphism of the lower parts of these piles to amphibolite, partial melting of the amphibolite to trondhjemitic-tonalitic liquids, and rise of these liquids — accompanied by continuing basaltic magmatism — to upper crustal or extrusive levels. Thus in a two-stage process siliceous rocks of low density are formed from the mantle. Trondhjemites thus occupy a key role in crustal genesis. Rocks of granitic composition ($K_2O \geqq 3\%$) may be generated in turn by anatexis of gray gneiss complexes or sediments derived therefrom, but must be considered of secondary importance in crustal genesis.

EXPERIMENTAL STUDIES OF THE EFFECTS OF VOLATILE COMPONENTS ON MELTING AND OTHER PHASE TRANSFORMATIONS IN THE TERRESTRIAL PLANETS, A. Boettcher, Department of Earth and Space Sciences, University of California, Los Angeles, Los Angeles, CA 90024.

To understand the origin and evolution of the terrestrial planets, it is necessary to understand the effects of volatile components on melting and other phase transformations.

At pressures below about 15 Kbars, H_2O is the most important of these components because of the large solubility in silicate liquids and the stabilities of refractory hydrous minerals, including amphiboles and micas. At higher pressures, CO_2 becomes very soluble in these liquids, and carbonates are more stable in peridotite than is CO_2-rich vapor. The effects of CH_4 and other reduced species are unknown, but they may be more stable than CO_2 + H_2O in the interiors of the Earth and other planets.

High-pressure experimental studies reveal that the *concentrations* of volatile species as well as the fugacities of H_2O, CO_2, H_2, O_2, and other components are required to shed light on endogeneous processes in the planets. Valuable estimates of these parameters can be gleaned from studies of the compositions of the atmospheres and crusts of these planets as well as from continued laboratory investigations and thermodynamic calculations, all of which are in infancy.

DEEP STRUCTURE OF THE CONTINENTAL CRUST FROM SEISMIC REFLECTION PROFILING, L. Brown, J. Brewer, F. Cook, L. Jensen, S. Kaufman, G. Long, J. Oliver, F. Schilt, Department of Geological Sciences, Cornell University, Ithaca, NY 14853

Structure in Archean basement rocks now at the surface can be exceedingly complex, with folds, intrusions, and faulting on several scales. This structural style is a reasonable analogue for basement rocks of the continental crust at depth. Traditionally applied geophysical methods have generally lacked sufficient resolution to map such complex structural trends at great depths. Recent efforts to apply high resolution seismic reflection techniques to the study of continental basement structure have significantly changed this situation. The most ambitious attempt is being guided by the Consortium for Continental Reflection Profiling (COCORP), using state-of-the-art multichannel seismic reflection methods. Over 40 sites have been proposed for COCORP field work, and surveys have been carried out at seven, representing a variety of geological problems in both the eastern and western U. S. The result is a large, internally consistent base of new, high resolution geophysical information bearing on both specific and general structural problems down to lower crustal and upper mantle depths.

Specific results include:

(a) Tracing the Laramide frontal fault of the Wind River Uplift to great depth as a moderate angle thrust, confirming compressional forces as its cause.
(b) Mapping an extensive, strong reflector at mid-crustal depths beneath the Rio Grande Rift near Socorro, New Mexico, corresponding to a magma body previously inferred on the basis of independent evidence.
(c) Imaging the San Andreas fault as a well defined, near vertical zone of disrupted reflection character, probably corresponding to a plane of intense plastic shearing.
(d) Demonstrating the [lateral] complexity [of] and lateral heterogeneity of the crust-mantle transition zone (Moho).

On a more general basis, these surveys establish the essentially heterogeneous nature of the crust, whose seismic character (and corresponding structural style) can be mapped in terms of:

(1) Continuous events corresponding to major faults, (meta-) sedimentary and igneous layering, or zones of partially molten material.
(2) Zones of short, discontinuous reflectors, possibly corresponding to deformed and intruded metamorphic terrains (Archean structural style?).
(3) Zones of transparent seismic character, most easily interpreted as homogenous (granitoid?) plutons.

Observation such as these provide a basis for mapping the distribution of metamorphic and igneous components in the crust, tracing major crustal faults to depth, identifying mechanisms of magma transport and emplacement, and investigating the role of deep structure in controlling the nature and location of shallow structures and resources.

While many interpretational problems remain, and the technique has proven more successful at addressing certain types of structural problems (e.g., thrust faulting) than others, it is clear that the results of reflection surveys are an important new tool for geologists and geophysicists attempting to decipher the architecture of the continental crust. Students of ancient crusts, in particular, should find that these new observations provide an improved perspective on the nature of crustal evolution.

EARLY ARCHEAN CRUSTAL RELATIONSHIPS IN THE SAGLEK-HEBRON AREA - NORTHERN LABRADOR

Kenneth D. Collerson[1], Dept. of Geology, Memorial Univ. of Newfoundland, St. John's, Nfld, CANADA.

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The Archean gneiss complex in the Saglek-Hebron area of northern Labrador has experienced a long and complex igneous, metamorphic and structural history which spanned circa 2000 Ma of Earth history. Despite this complexity it has been possible by careful observation and field mapping to establish a relative chronology (Table 1) between the different major lithological units (Bridgwater et al., 1975; Collerson et al., 1976). This relative chronology of events is in part based on the relationship between a swarm of characteristic meta-diabase dykes (Saglek dykes) and the different gneissic lithologies, and on rarely observed intrusive relationships among the igneous protoliths of some of the gneisses, preserved in domains of low finite strain. The stratigraphy of the gneiss terrain is broadly subdivided into: (1) lithologies which pre-date the dykes and, (2) lithologies formed from protoliths emplaced, laid down or tecttonically refoliated after intrusion of the dykes. It is comparable to the sequence of events recognised by McGregor (1973) for the Archean gneiss complex in the Godthaab district of West Greenland.

In Northern Labrador the early Archean crust (pre-Saglek dyke) is composed predominantly of either migmatitic or porphyroclastic (augen) quartzo-feldspathic gneisses which contain inclusions of earlier supracrustal lithologies (the Nulliak assemblage) and layered gabbros and anorthosites (the Mentzel Intrusive Association).

TABLE 1: MAJOR EVENTS IN THE ARCHEAN GNEISS COMPLEX - SAGLEK

	Pre-Saglek dyke events.	
1.	Early crust	
2.	Deposition and extrusion of the Nulliak assemblage	
3.	High-grade metamorphism	
4.	Emplacement of the protoliths of the Uivak I gneisses	C.3800 Ma.
5.	Deformation - metamorphism of the Uivak I gneisses	C.3800 Ma.
6.	Emplacement of the protoliths of the Mentzel Intrusive Association and the Uivak II gneisses	
7.	Deformation - metamorphism, formation of Uivak II gneisses	C.3600 Ma.
8.	Intrusion of the Saglek dykes	
	Post-Saglek dyke events.	
9.	Deformation	3420 Ma.
10.	Deposition and extrusion of the Upernavik supracrustals?	
11.	Emplacement of layered intrusions of ultramafic, gabbroic and anorthositic composition.	
12.	Intercalation Uivak gneisses and the Upernavik supracrustals.	Pre 3000 Ma.
13.	Emplacement of syntectonic tonalitic and granitic sheets.	C. 2800 Ma.
14.	Reactivation of the gneiss complex formation of Kiyuktek gneiss (13-14 penecontemporaneous associated with granulite - amphibolite facies metamorphism)	C.2700 Ma.
15.	Emplacement post-tectonic granites	2500 Ma.
16.	Emplacement diabase dykes	C.1900 Ma.
17.	Weak metamorphism (hudsonian)	1750 Ma.

Outcrops of the Nulliak assemblage range from semi-continuous units 2-3 km along strike and 50-100 m wide to tectonically fragmented, highly attenuated inclusions seldom larger than 1 x 0.25 m in size. The Nulliak assemblage includes banded and homogeneous mafic to felsic lithologies which are thought to be derived from a basaltic to felsic volcanic sequence. Finely layered Mg-rich clinopyroxene-hornblendites, associated with the meta-volcanics, are regarded either as parts of differentiated basic igneous intrusions (pyroxenites) or as komatiitic extrusives. Sediments within the Nulliak assemblage include banded silicate-oxide iron formation, carbonates and minor semi pelites. Although the dominant metamorphic grade of the gneiss complex is amphibolite facies some exposures of Nulliak lithologies exhibit cores with

EARLY ARCHEAN CRUST IN LABRADOR

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granulite facies mineralogy.

The migmatitic and augen gneisses are subdivided into two suites (the Uivak I and Uivak II gneisses). The Uivak I gneisses which comprise over 80% of the exposed early Archean rocks in northern Labrador, are a migmatitic to relatively homogeneous, dominantly leucocratic, medium to fine grained suite of grey gneisses. They commonly contain prominent mesoscopic and microscopic veins of K-fledspar rich pegmatite which predate the intrusion of the Uivak II gneisses and the Saglek dykes. Homogeneous portions of the gneisses are dominated by quartz-oligoclase-microcline-biotite-hornblende-bearing trondhjemitic, tonalitic and granodioritic compositions interleaved with septa of more leucocratic gneiss. Sheets of melanocratic hornblende or biotite-rich gabbroic, dioritic and monzonitic gneiss up to 1 m wide are present in subordinate amounts and represent more mafic phases of the Uivak I suite. Geochemical and microstructural studies have shown that the Uivak I suite formed mainly from trondhjemitic to granodioritic parents, which were deformed and metamorphosed prior to the emplacement of the Uivak II gneisses. Isotopic studies in progress at A.N.U. indicate that the Uivak I gneisses may be as old as 3850 Ma (Fig. 1).

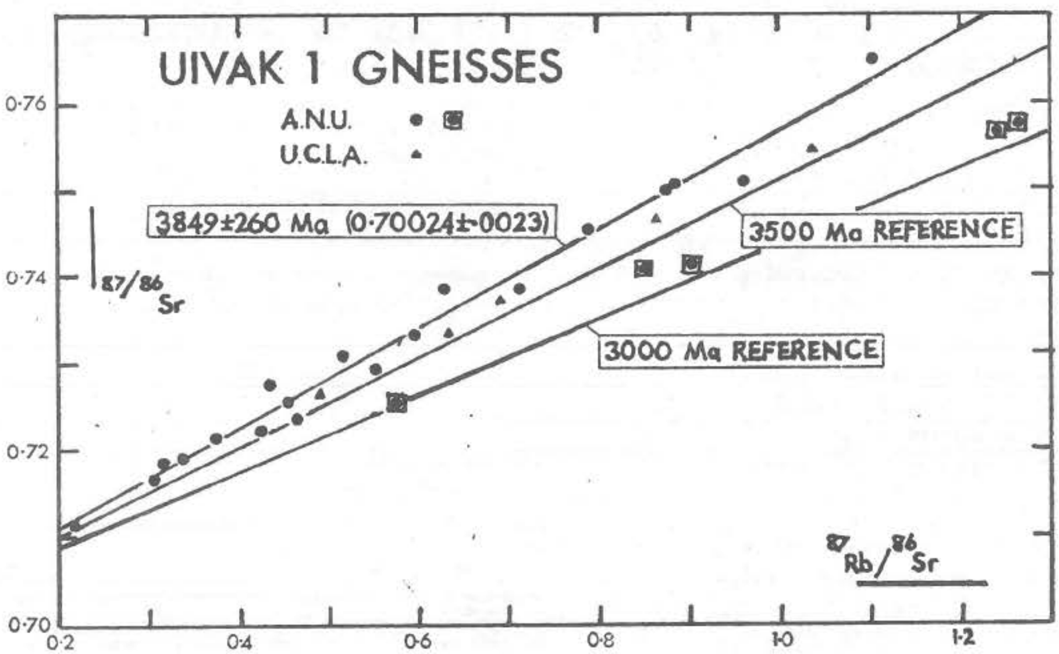

Fig. 1. Rb-Sr evolution diagram for the Uivak I gneisses($^{87}Rb\ \lambda=1.42\times10^{-11}y^{-1}$).

The Uivak II augen gneisses are a medium to coarse grained granitic to granodioritic suite. They are more potassic and are richer in ferromagnesian minerals than Uivak I gneisses with equivalent SiO_2 content. In terms of areal abundance they constitute less than 10% of the gneiss complex. In areas of relatively low strain intrusive igneous relationships with the Nulliak assemblage supracrustal lithologies and with the Uivak I suite are preserved. In these areas outcrop widths range up to 1 km. In other areas the Uivak II gneisses form narrow sheets 4 to 5 in width with intrusive contacts against Uivak I migmatites.

Mega-xenoliths (up to 20 x 5 m in size) of a disrupted compositionally layered meta-gabbro-leucogabbro-anorthosite-diorite plutonic body (the Mentzel Intrusive Association,occur in the Uivak II gneisses at a number of

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localities and are interpreted as representing an early phase of this suite.

The early metamorphic history of the terrain involved the formation of both granulite and amphibolite facies assemblages at different structural levels. Microstructural and paragenetic relationships in the Uivak I grey gneiss indicates that retrogression from granulite to amphibolite facies occurred after the establishment of the planar anisotropy in the suite. This regression involved the development of biotite and secondary hornblende from hornblende and orthopyroxene in the early granulite facies gneisses. It was also associated with an increase in model microcline and quartz. The alteration in the suite is interpreted as being the result of K and Rb metasomatism possibly associated with the emplacement of deformation of the parents of the Uivak II gneisses. The increase in Rb produced lower K/Rb and higher Rb/Sr ratios than are normally observed in trondhjemitic-tonalitic associations.

Highly fractionated REE patterns and variable Ce/Y in the Uivak I gneisses are interpreted to be induced either by metamorphic or igneous processes. For example, these processes may have involved the transport of heavy REE's and Y (as well as other LIL elements) in the form of soluble complexes with volatiles escaping from the lower crust or upper mantle. An alternative explanation is that the patterns were induced by removal of refractory phases during the partial melting events which derived the igneous protoliths of the Uivak I gneisses.

In Collerson and Bridgwater (1979) a model is presented whereby the trondhjemitic and tonalitic parents of the Uivak I gneisses are generated by partial melting of mafic granulite under the influence of a high, early Archean geothermal gradient associated with massive volatile outgassing from the mantle and lower crust. The production of early Archean sialic crust is envisaged as a cumulative, cyclic process, mafic and ultramafic volcanism occurring synchronously with the partial melting at depth of earlier formed mafic granulite facies crust. The Uivak II gneisses are considered to have formed by partial melting of granulite facies tonalitic and trondhjemitic gneisses, possibly with some fractional crystallization of the derived melt.

A summary of the geological evolution of the complex after the emplacement of the Saglek dykes is also presented in Table 1. Elaboration is however beyond the scope of this presentation and details have been described more fully elsewhere (Collerson et al., 1976; Bridgwater et al., 1978).

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THE INTERPRETATION OF SCATTERED Rb-Sr ISOTOPIC DATA FOR EARLY ARCHEAN GNEISSES FROM LABRADOR AND GREENLAND.

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and

Kenneth D. Collerson. Department of Geology, Memorial University, St. John's, Nfld. Canada.

Rb-Sr isotopic data for both the Amitsoq grey gneisses in Greenland and the Uivak I gneisses in Labrador display a scatter which exceeds that predicted by experimental error. (Figure 1). Perfect isochrons are therefore not defined presumably as a result of geological effects. Moorbath et al., (1972) in interpreting results for the Amitsoq gneisses, although acknowledging the imperfect fit for the West Greenland data, did not explicitly model these geological effects. From an analysis of the 1972 data it appears that a McIntyre et al. (1966) Model 3 solution has been used.

In this, all the geological error is attributed to a real variation in initial $^{87}Sr/^{86}Sr$ between samples which is independent of their $^{87}Rb/^{86}Sr$. The samples are treated as completely closed to Rb,Sr movement since their common age at ∿3600 Ma, despite the manifest geological evidence for several later metamorphic events.

Such an interpretation cannot be correct in general, because the range indicated for the population of initial $^{87}Sr/^{86}Sr$ from the Praestefjord area extends well into the forbidden region below BABI (Fig. 2). Values below BABI are also displayed by Amitsoq gneisses from the Narssaq, Qilagerssuit area (Baadsgaard et al., 1976) and the Uivak I gneisses (Fig. 2).

To avoid this dilemma, we propose here a new regression method that is based on a more realistic assessment of the behaviour of Rb-Sr total-rock systems. It is well known that biotites and other minerals in the Amitsoq gneiss were open systems at approximately 1600 to 1800 Ma (Pankhurst et al., 1973, Baadsgaard et al., 1976), which corresponds in time to a period of major Proterozoic crustal modification (Hudsonian Orogeny). In addition, adjacent migmatite bands from the Uivak gneisses, as total rock samples, also show open system response to metamorphism at this time (Fig. 3). We therefore propose that the geological scatter of the larger and more homogeneous Amitsoq and Uivak gneisses should also be attributed to open-system behaviour at ∿1750 (and earlier) rather than real variations in the initial $^{87}Sr/^{86}Sr$. In particular, we propose a model of local Sr isotope homogenization between adjacent total rock samples, but with preservation of regional differences in $^{87}Sr/^{86}Sr$ and $^{87}Rb/^{86}Sr$. Fig. 4 depicts the isotopic patterns immediately prior to the 1750 Ma event, immediately after it, and at the present day.

A simple regression method can be applied if the measured data are first transformed to the immediately-after 1750 Ma situation. This can be done by subtracting $(e^{\lambda t}-1)$ $^{87}Rb/^{86}Sr$ for each sample from the corresponding $^{87}Sr/^{86}Sr$, where t is 1750 Ma and λ the ^{87}Rb decay constant. Inspection of Fig. 4(b) shows that the estimation of the original pre-1750 Ma isochron requires the prediction of the regional mean $^{87}Rb/^{86}Sr$ that correspond to the various $^{87}Sr/^{86}Sr$ at 1750 Ma. (In other words, it is the regression of x against y rather than y against x as employed by McIntyre and York).

An independent and optionally-extra step in the data analysis is a procedure for constraining the mean initial $^{87}Sr/^{86}Sr$ estimated for the

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protolith rocks to realistic values. The mean initial $^{87}Sr/^{86}Sr$ can be modelled on single-stage ^{87}Sr evolution using the "whole-earth" parameters proposed by the Cal. Tech and Lamont-Doherty laboratories by comparative Rb-Sr and Sm-Nd work ($^{87}Rb/^{86}Sr$ 0.081, present-day $^{87}Sr/^{86}Sr$ 0.7045). The regression line at 1750 Ma may then be forced to pass through the 1750 Ma whole-earth parameters (e.g. Acton 1959, p. 17,18), which ensures that the estimates of age and initial $^{87}Sr/^{86}Sr$ are mutually consistent.

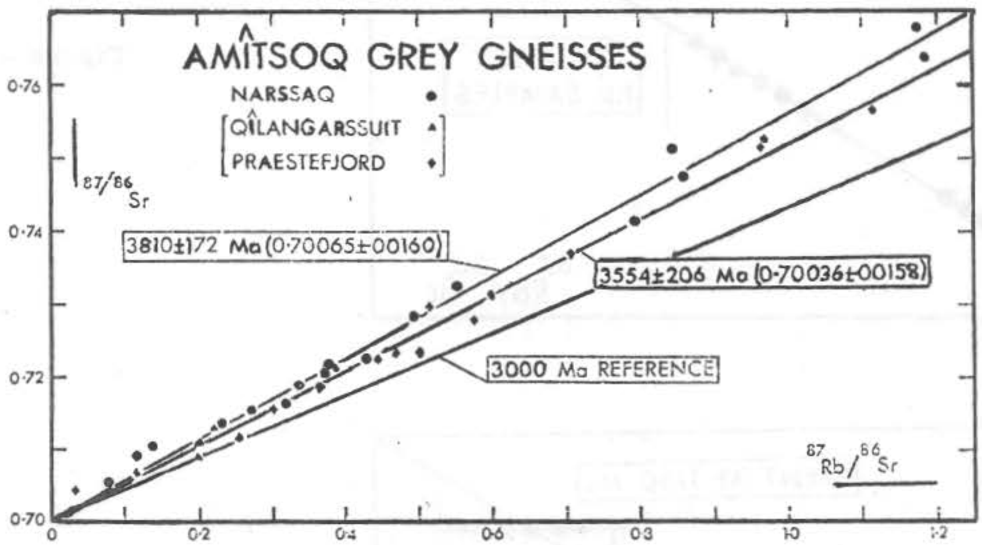

Figure 1

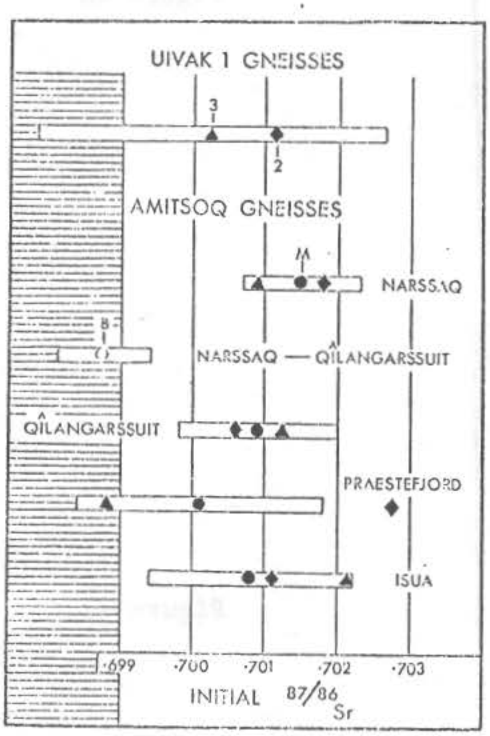

Figure 2.

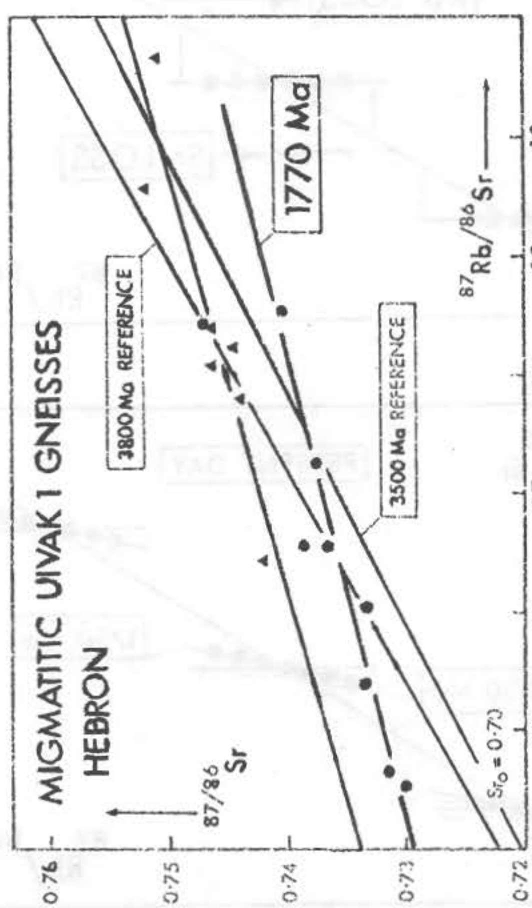

Figure 3

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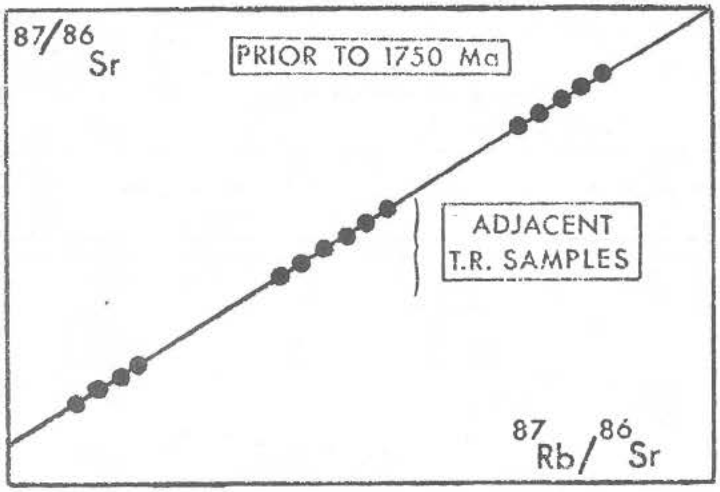

Figure 4a

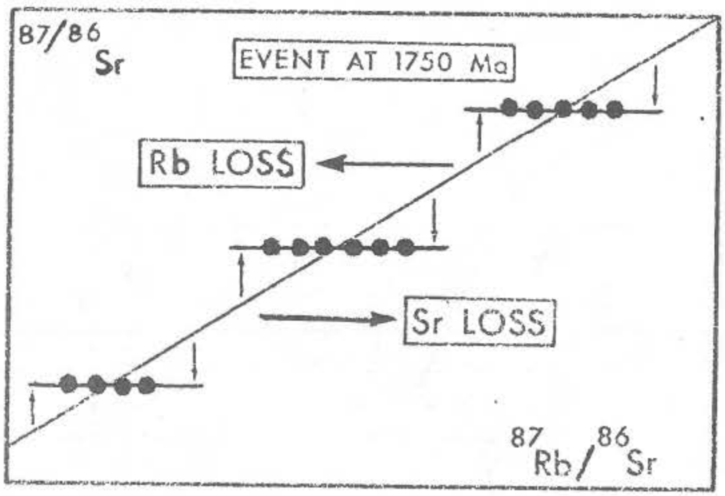

Figure 4b

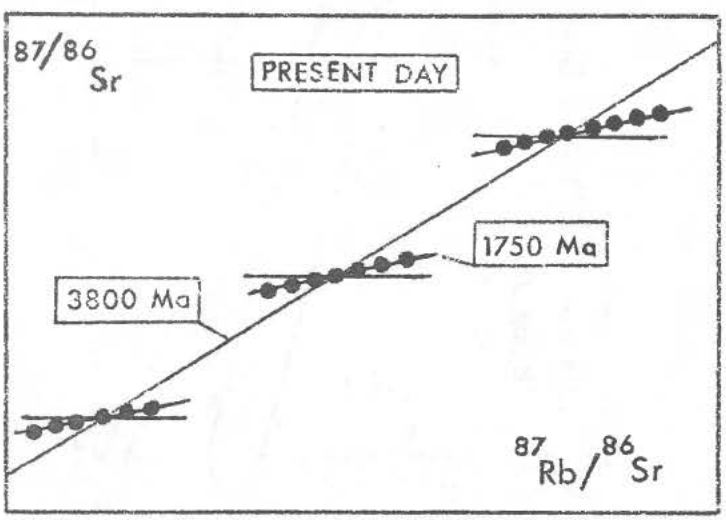

Figure 4c

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HYDROUS MINERALS IN EARTH'S UPPER MANTLE: STORAGE OF ALKALI METALS AND HALOGENS: SPECULATIONS ON VENUS AND MARS; J.S. Delaney, R.L. Hervig, J.V. Smith, Dept. of the Geophysical Sciences, University of Chicago, Chicago, IL, 60637; J.B. Dawson, D.A. Carswell, Dept. of Geology, University of Sheffield, Sheffield, England.

The present storage of volatile elements in the upper mantle provides clues to geochemical models for these elements, and allows speculations for Venus and Mars. Unquestionably the volatile elements are responsible for extensive metasomatism, but we believe that careful petrographic study combined with electron microprobe analysis has revealed mineral hosts in metamorphic equilibrium with silicate minerals.

Hydrous phases in the upper mantle. Phlogopite mica and amphibole are present in many peridotite and eclogite xenoliths brought up by kimberlite magmas from the upper mantle. Interpretation of petrographic textures is difficult. Undoubtedly there is pervasive metasomatism, perhaps associated with melting events, but some micas are in textural equilibrium with anhydrous silicates (1). Table 1 lists the principal types of mica and amphibole found in peridotite xenoliths. Generalized stability curves for mantle phlogopite and amphibole are given in Fig. 2 of the abstract by Smith, Newton and Janardhan. Typical geotherms result in stability limits of 60 to 100km depth for amphibole and ~150km for phlogopite. Groundmass micas from kimberlite are described in (4). We emphasize the chemical composition of micas (Table 1, col. 1) which occur in large grains (~0.5-2mm across) in textural equilibrium with anhydrous silicates of comparable grain size in garnet lherzolites with granular texture. These micas have similar composition irrespective of the location of the kimberlite pipe in southern Africa. Those with higher Ti occur coarsely intergrown with diopside in a texture suggestive of simultaneous crystallization from residual liquid. Secondary micas have different compositions depending whether they replace garnet or orthopyroxene. Mica megacrysts in kimberlite magma are richer in Fe and lower in Cr and Ni. Both types of amphiboles carry substantial Na and K.

K,Rb and Ba in micas from kimberlite and peridotitic xenoliths, and implications for origin of basaltic rocks. This section is a brief abstract of (5). Special electron microprobe analyses of Rb and Ba in micas (~40ppmw,2σ) were plotted on a diagram of log K/Ba vs. log K/Rb. Both primary- and secondary-textured micas from peridotite have a negative correlation between K/Ba and K/Rb. This is surprising in view of data for biotites crystallized near the Earth's surface. The negative correlation for high-pressure phlogopites may result from partial melting in the mantle, or perhaps from a mica-clinopyroxene exchange reaction.

A model composition for the bulk Earth (based on same K/Rb as C1 meteorite and one-sixth K/Ba) falls in the range for peridotite micas. Most volcanic rocks fall on a single positive trend between K/Rb and K/Ba, which apparently poses a problem for differentiation models based just on mica-liquid partitioning. Most continental varieties have K/Rb (~300) and K/Ba (~30) which lie inside the range for peridotitic micas and close to the model values for the bulk Earth. Tholeiite and alkali basalt from oceanic islands have higher K/Rb (430), and most mid-ocean ridge basalts have much higher K/Rb (~1100). Whereas most continental volcanic rocks could obtain their K,Rb and Ba simply from total extraction of phlogopite during partial melting of peridotite, rocks with high K/Rb could not, unless the phlogopite had a different composition than in xenoliths from kimberlite pipes. Data for amphibole and clinopyroxene do not allow simple models for total extraction of K,Rb and Ba. Complex models involving partial extraction of

HYDROUS MINERALS IN EARTH'S UPPER MANTLE:

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Table 1. Electron microprobe analyses of hydrous minerals from upper mantle.

	1	2	3	4	5	6
N	17	9	5	22	1	22
SiO_2	40.95(1.06)	39.4(0.9)	41.2(1.7)	41.0(1.4)	45.1	55.4
TiO_2	0.18(0.16)	0.9(0.7)	0.25(0.23)	1.22(0.74)	0.0	0.46
Al_2O_3	13.45(0.68)	15.4(2.2)	13.0(1.8)	11.9(1.2)	11.1	0.99
Cr_2O_3	0.84(0.10)	1.7(0.8)	0.88(0.37)	0.40(0.36)	2.20	0.48
FeO	2.60(0.33)	3.4(0.6)	2.8(1.0)	4.8(2.3)	2.69	2.25
MnO	0.02(0.01)	0.04(0.04)	0.03(0.03)	0.03(0.03)	0.18	-
MgO	26.05(0.78)	24.0(1.5)	26.2(2.3)	23.4(2.2)	20.3	22.9
NiO	0.22(0.03)	0.13(0.06)	0.21(0.12)	0.11(0.07)	0.36	-
CaO	0.01(0.01)	0.02(0.02)	0.02(0.02)	0.01(0.01)	10.9	6.74
Na_2O	0.35(0.22)	0.72(0.35)	0.41(0.16)	0.27(0.26)	3.23	3.82
K_2O	9.98(0.69)	9.2(0.5)	9.2(0.7)	10.7(0.7)	1.34	4.63
Cl	0.12(0.12)	0.04(0.02)	0.02(0.01)	0.12(0.22)	-	-
F	0.51(0.24)	0.52(0.27)	0.34(0.14)	0.82(0.15)	-	-
BaO	0.29(0.20)	0.18(0.13)	0.24(0.07)	0.04(0.03)	-	-
Rb_2O	0.027(0.008)	0.02(0.01)	0.02(0.02)	0.08(0.02)	-	-
Sum	95.597	95.67	94.82	94.90	97.40	97.67

1-3 micas in peridotite xenoliths from kimberlite; 1 primary texture, 2 replacing garnet, 3 replacing orthopyroxene; 4 mica megacrysts in kimberlite; 5 chromic pargasitic hornblende in harzburgite xenolith (2); 6 metasomatic richterites (3). S.D. in brackets. N number of samples.

alkalies, combination of minerals, and crystal-liquid fractionation during ascent may be needed.

Bulk kimberlite and carbonatite, and inclusions in diamond, have much lower K/Rb and K/Ba than peridotitic micas, but the reason(s) for these low ratios are unknown. If these rock types result from melting deeper than 150km, the low ratios may be giving clues to storage of alkalies beneath the uppermost mantle. At high pressure, garnet stores Na and clinopyroxene stores K (6,7), and the hollandite and perovskite structure types are potential hosts for large cations (8).

F and Cl in mica, amphibole and apatite, and implications for origin of basaltic rocks. So far, F and Cl have been detected in xenoliths of upper-mantle lherzolite and harzburgite xenoliths only in the minerals phlogopite, amphibole and apatite. Typically the micas have high F/Cl (Table 1), which feature is also found for amphibole (F 0.5-1 wt.%; Cl <30-300ppmw). No analyses of primary apatite have been made, but secondary apatite from lherzolite nodules in kimberlite contain ~3 wt.% F, while a megacryst from the Moses Rock "kimberlite" contains ~0.5% F and <100ppmw Cl. Thus all three types of minerals have high F/Cl, and no Cl-rich mineral has yet been found in a nodule of mantle origin.

Metasomatism from host kimberlite magma is not responsible for the differences between the Cl and F contents of micas in peridotites (Table 1), based on simple comparison of chemical compositions.

Although the halogen content of the hydrous minerals (strictly speaking, hydroxylated minerals) should be characteristic therefore of the upper mantle, perhaps other halogen-bearing minerals may be unrecognized.

Table 2 lists data pertinent to the possible origin of basalts from partial melting of peridotite in the upper mantle. Any conclusions are highly tentative at this stage, because of incomplete data for Cl and F in basalts and differentiated rocks, and because of absence of analyses for apatite in textural equilibrium with silicates in peridotite xenoliths. Bulk analyses in (13) of peridotite xenoliths and mica megacrysts in kimberlite

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Table 2. K,F and Cl in hydroxylated minerals, basalts and C1 chondrite.

wt.%	K	F	Cl	K/F	K/Cl	F/Cl
primary mica, peridotite	8.28	0.51	0.12	9-52	78-220	2-22
secondary mica, peridotite	7.64	0.46	0.03	8-30	87-1400	4-114
megacryst mica, kimberlite	8.88	0.82	0.12	8-16	15-500	2-45
MARID amphibole	4.1	0.80	0.008	4-7	>130	25-160
apatite, Moses Rock kimberlite	-	0.45	0.007	-	-	64
apatite, serpentine vein, peridotite	-	~3	<0.010	-	-	>300
basalts, Iceland (9)	0.55	-	0.024	-	2-60	-
apatite, Shonkin Sag (10)	-	2.6	0.33	-	-	7-9
apatite, Skaergaard (11)	-	2.72	0.43	-	-	2-40
C1 chondrite (12)	0.054	0.030	0.05	1.5	0.89	0.6
kimberlite (13)	-	0.16	0.015	-	-	11
basalts (quoted in 13)	-	0.04	0.006	-	-	7
megacryst mica, kimberlite (13)	-	0.45	0.045	-	-	10
sheared nodules, kimberlite (13)	-	0.018	0.015	-	-	1.2
unsheared nodules, kimberlite (13)	-	0.049	0.002	-	-	24

match our ranges of individual micas, but the ranges are rather wide. Further study is needed to determine whether all unsheared peridotite nodules have higher F/Cl than sheared ones, as for the few specimens listed in (13). Naive comparison of F/Cl ratios of basalts quoted in (13) with those of our peridotitic micas suggests that mica ± minor apatite could supply the F and Cl in the basalt magmas by partial melting of peridotite. For Icelandic basalts (9), the high Cl and H_2O content were thought to require a mantle source rich in Cl and H_2O. Might this source be related to sheared peridotites? The micas and apatites in Table 2 have lower Cl/F than for C1 chondrite.

Storage of alkali metals and halogens in Venus and Mars. Controls on the stability of amphibole and mica are given in Fig. 2 of the abstract by Smith, Newton and Janardhan. Of course, if Venus has lost all its water, mica and amphibole would not be stable, except perhaps for oxy-varieties. By analogy with terrestrial peridotites, K might occur in mantle clinopyroxene in Venus, and perhaps even in a high-pressure structure. Feldspar should be stable near the surface in the range 0-~10kb (cf. Figs. 2 and 8 of ref. 14) and is probably the major host of K,Rb and Cs. Although the hollandite polymorph of K-feldspar should be stable to great depth, K may have been flushed to the surface during crystal-liquid differentiation. For Mars, there are several possible hosts for the alkalies, of which certain evaporite minerals, feldspar, amphibole and mica are likely to be the most important. Storage of halogens inside of Venus is feasible with apatite, fluorite, mica and amphibole providing potential reservoirs for F, and scapolite, cancrinite and apatite for Cl. If active volcanoes are located on Venus, a fly-over mission to sample escaping gas would provide useful clues to the storage of volatile elements in the interior. Numerous possibilities are available for Mars, including low-T minerals as well as the ones listed for Venus.

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PETROLOGIC HETEROGENEITY IN THE UPPER MANTLE OF THE EARTH: BARREN AND FERTILE HARZBURGITES: MANTLE SECTION AT MALAITA, SOLOMON ISLANDS: IMPLICATIONS FOR VENUS AND MARS; J.S. Delaney, R.L. Hervig, J.V. Smith, Dept. of the Geophysical Sciences, University of Chicago, Chicago, IL 60637; J.B. Dawson, Dept. of Geology, University of Sheffield, Sheffield, England; P.H. Nixon, Dept. of Geology, University of Papua New Guinea.

Partial melting of lherzolite to give basaltic magma and a harzburgite residue is an accepted mechanism for differentiation of the Earth's upper mantle. We report electron and ion microprobe analyses of minerals in peridotite xenoliths in S. African kimberlite and Malaita alnöite.

Mantle section at Malaita, Solomon Islands. Preliminary studies of megacrysts and lherzolite xenoliths in the alnöite (1,2) were followed by detailed interpretation of the mineral chemistry. Fig. 1 shows pressure and temperature estimates from coexisting pyroxenes in two types of garnet lherzolite xenoliths, and for orthopyroxene and clinopyroxene megacrysts interpreted to coexist because of minor-element correlations. Such P,T estimates are controversial, but taken at face value, the mantle below Malaita can be interpreted in terms of the section in Fig. 2. Below the Moho a transition occurs from lherzolites carrying Al-rich spinel (C) to garnet lherzolites ranging from type B to type A with increasing depth. Local increase of temperature (Fig. 1) caused partial melting of lherzolite. The magma underwent crystal-liquid differentiation in one or more magma chambers, and the differentiation was interpreted from trends in the megacrysts of Fe/Mg,Na,Ti, and Cr. The most Mg-rich megacrysts are garnet, diopside and bronzite. These are followed by lamellar intergrowths of ilmenite and sub-calcic diopside, and finally by augite. Alnöitic magma from greater depth ripped off samples of the lherzolites and the magmatic differentiates.

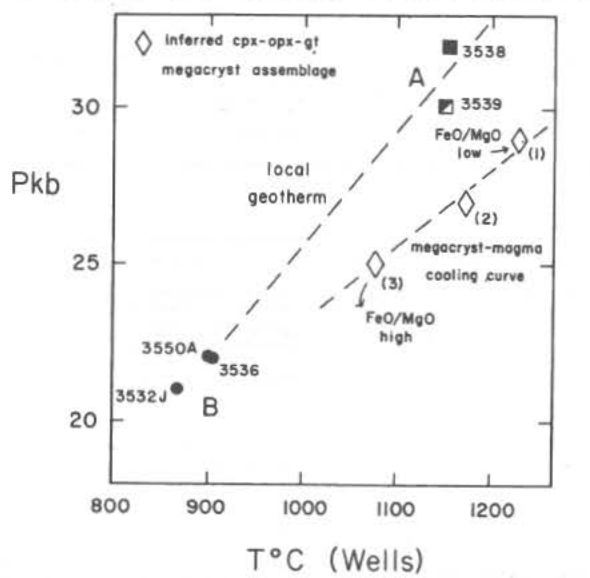

Fig. 1. Estimated P and T for coexisting pyroxenes (Wells thermometer; Newton barometer) from garnet lherzolite xenoliths (A high-Cr; B low-Cr) and from garnet, bronzite and diopside megacrysts interpreted to coexist at three stages of crystal-liquid differentiation from low Fe/Mg to high Fe/Mg.

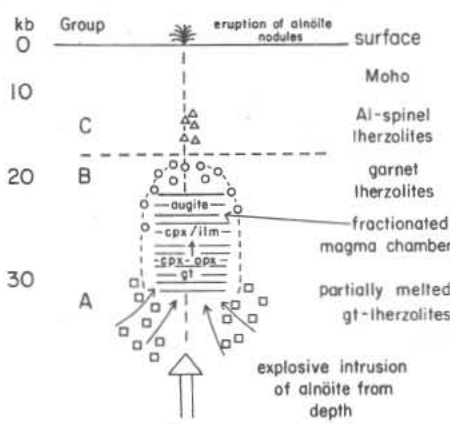

Fig. 2. Cartoon of upper mantle under Malaita showing A and B types of garnet lherzolites and overlying C type of spinel lherzolites. The magma chamber is purely schematic, and it is exaggerated at least 10-fold horizontally. A series of magma chambers, occupied successively may be more likely. There is no depth control on the augite and cpx/ilm intergrowths, and they may have crystallized in magma chambers in the sp-lhz field.

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Mineral chemistry in harzburgite and lherzolite xenoliths from S. Africa and western USA (3). Harzburgite xenoliths can be separated into barren and fertile groups from the low or high content, respectively, of Ca,Al and Cr in the orthopyroxene (e.g. Fig. 3). Although overlap occurs in Fig. 3 for garnet and spinel lherzolites and harzburgites, all four groups separate in multidimensional chemical space when several minor elements in both opx and olivine are considered. Ion microprobe analyses of trace elements in olivine (Fig. 4) show a general trend for Li,Na,P and Ti to decrease from lherzolite and dunite hosts to harzburgite. Surprisingly the concentrations of these magmaphile elements tend to be higher in olivines from barren than from fertile harzburgites. Ion probe analyses of opx are in progress.

Melting-metamorphism relationship between harzburgites and lherzolites (3). The simplest assumption in upper-mantle models, that garnet lherzolite is a relatively homogeneous rock capable of differentiation into a basaltic liquid and a "barren" harzburgite residuum, must be modified to take into account different degrees of partial melting and subsequent cooling to metamorphic assemblages (4). Rare opx megacrysts (e.g. Table 1, col. 1) contain substantial Al_2O_3, Cr_2O_3 and CaO, and some are exsolving garnet and diopside in kink bands (3). Mathematically, opx from megacrysts and fertile harzburgite (col. 2) could be reconstituted into a mixture of garnet, clinopyroxene, and opx like that in barren harzburgite. Indeed many garnet lherzolites contain <5% garnet and <2% cpx, and might be the metamorphosed product of fertile harzburgite (e.g. compare cols. 1 and 2 with col. 5 in Table 5; the Na_2O content is discussed in ref. 3). Furthermore partial melting with release of basalt magma might convert a fertile into a barren harzburgite. Some ways for generating mantle heterogeneity by melting of lherzolites and harzburgites to form basaltic magmas, residues and cumulates are listed in the flow sheet. Retrograde metamorphism reduces solid solution, thereby converting an ol-opx rock (harzburgite) into a depleted lherzolite (mainly ol+opx; minor gt+cpx). Melting may be single- or multi-stage.

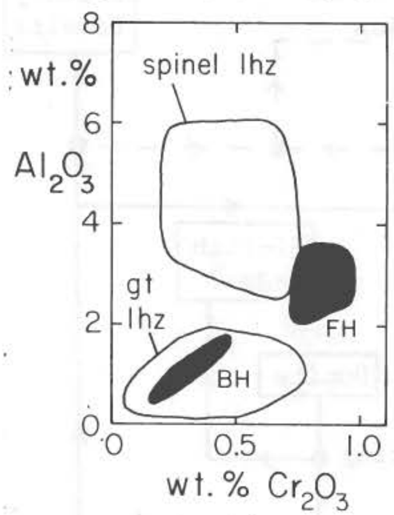

Fig. 3. Al_2O_3 and Cr_2O_3 fields for orthopyroxenes from barren and fertile harzburgites and garnet and spinel.

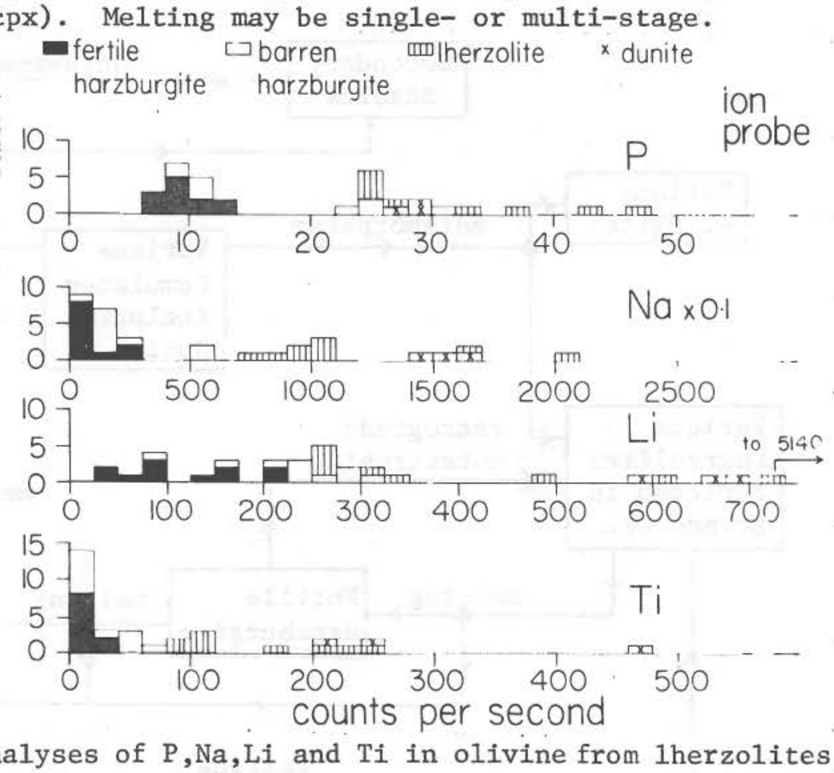

Fig. 4. Ion-microprobe analyses of P,Na,Li and Ti in olivine from lherzolites, dunites, and barren and fertile harzburgites.

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Table 1. Chemical composition of opx.

	1	2	3	4	5
SiO_2	55.2	57.4	58.5	58.1	56.4
TiO_2	tr.	0.00	0.00	0.06	0.07
Al_2O_3	3.88	2.91	0.96	0.82	2.53
Cr_2O_3	0.76	0.79	0.32	0.32	0.74
Fe_2O_3	1.20	-	-	-	-
FeO	3.68	4.43	4.24	4.80	5.09
MnO	0.14	0.06	0.10	0.19	0.12
MgO	33.1	34.9	35.9	35.5	33.9
NiO	n.a.	0.08	0.07	0.11	0.10
CaO	1.97	0.83	0.24	0.46	0.97
Na_2O	0.26	0.03	0.02	0.12	0.12
Sum	100.19	100.47	100.35	100.39	100

1. mean of two megacrysts veined by garnet and diopside, Frank Smith mine. 2,3 mean orthopyroxene from 11 fertile and 13 barren harzburgite xenoliths; two fertile harzburgites have opx with Na_2O 0.16%. 4. mean orthopyroxene from 14 garnet lherzolite xenoliths. 5. theoretical opx calculated from 0.7% Cr-diopside, 9% garnet and 0.3% Cr-spinel and 90% opx with composition typical for garnet lherzolite (3).

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Application to Venus and Mars. Using a new heterogeneous accretion model (5), Venus could be similar to the Earth except for possible loss of all water. Partial melting and retrograde metamorphism could lead to a complex upper mantle with various lherzolites, harzburgites and eclogites. The Martian mantle could also be peridotitic (6) and massive extrusion of Fe-rich basalts would have allowed ample scope for development of a complex upper mantle. Particularly important is whether harzburgites will have floated over lherzolites, and what is the effect on the moment-of-inertia.

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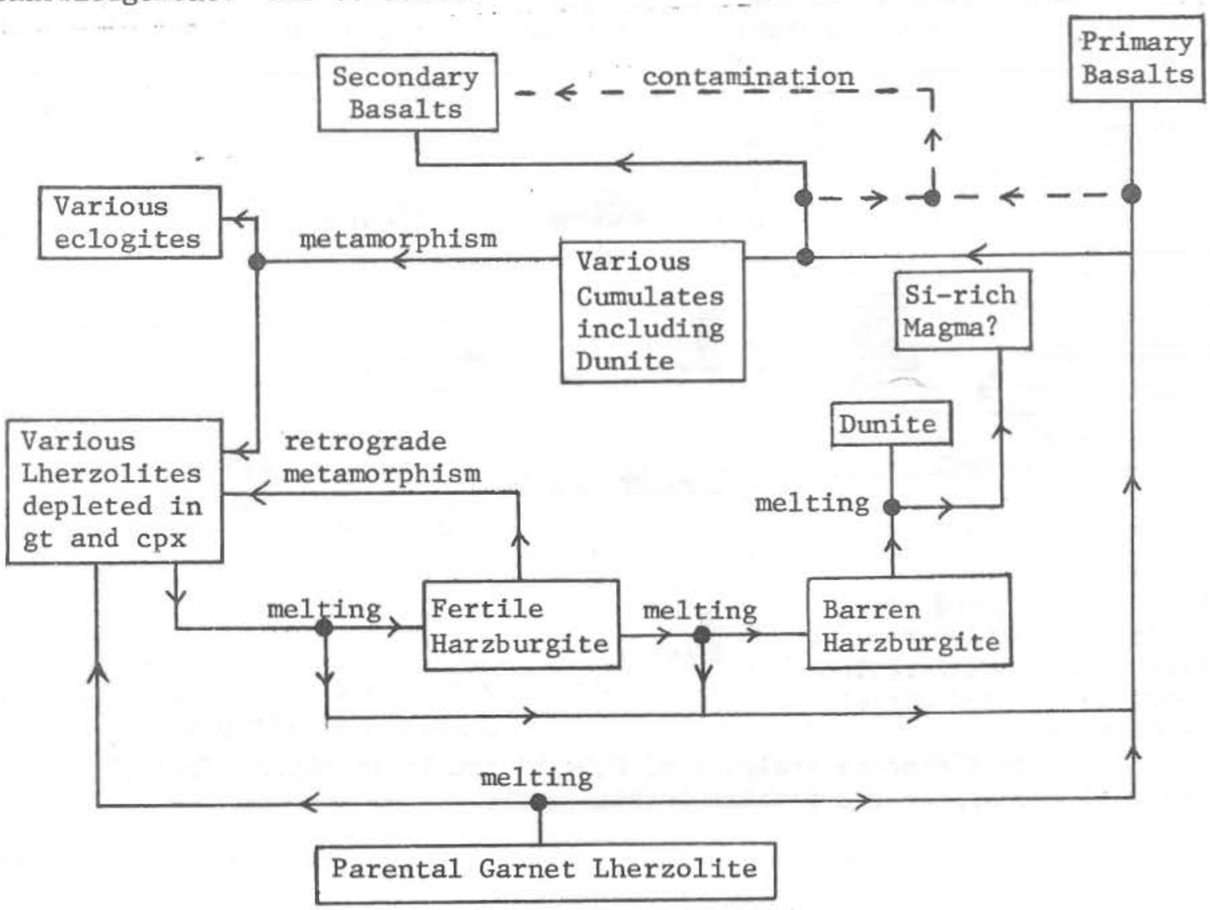

AGE RELATIONSHIPS IN ARCHEAN ROCKS; S.S. Goldich, Department of Geology, Colorado School of Mines, Golden, CO 80401 and U.S. Geological Survey, Box 25046, Federal Center, Denver, CO 80225

With the adoption of the two-fold division of the Precambrian, the time boundary between the Archean and Proterozoic has been set at 2,500 m.y. Ages, ranging from 3,300 to 3,800 m.y., have been fairly well established on all the continents. West Greenland has large, well-exposed areas of ancient rocks that, in part, have escaped drastic changes in younger geologic time. Other parts of the world present formidable challenges to geochronologists because of complex structure, multiple igneous activity, polymetamorphism, and commonly poor exposures, but continuing efforts have yielded results and have advanced our understanding of geologic history and processes. In the United States, more or less coordinated geochronological and geological studies are continuing in Michigan, Wisconsin, Minnesota, Montana, Wyoming, and Utah. Significant developments are the 3,000-m.y. isotopic ages and the abundant petrographic and chemical data which show that the most ancient rocks are similar mineralogically and chemically to very young rocks.

THE POTENTIAL EFFECTS OF METEORITE BOMBARDMENT ON THE EVOLUTION OF THE PROTO-CRUST OF THE EARTH. Richard A.F. Grieve, Earth Physics Branch, E.M.R., Ottawa, Canada, K1A 0Y3

The crater density distribution on Mars, Mercury and the moon indicates that they underwent a period when surface evolution was heavily influenced by impact processes. On the moon, a high flux of impacting bodies, lasting till ∿3.9 b.y., resulted in the formation of the major multi-ring basins, which were later filled to varying degrees by mare basalts. The earth has no obvious record of this intense bombardment but many models for the early earth assume it was subjected to the same high flux (1,2). Dynamical arguments (3) and the similarity between the Phanerozoic cratering rate on earth and the post-mare (<3.0 b.y.) rate on the moon (4) indicate that this is a valid assumption. However, problems arise in interpreting the effects of this bombardment, which has been called upon to produce such widely divergent results as the initial ocean basins (5) and the locus of initial sialic crust generation (6). This divergence results in part from the lack of a consensus on the nature of the proto-crust of the earth and the use of relatively direct analogies with the response of the early lunar crust to bombardment. For discussion, it is assumed here that earth had a thin, <15 km, world-wide differentiated crust of unspecified composition during 4.6-3.9 b.y. If a stable crust was not developed till after 3.9 b.y. (7) much of the following is irrelevant.

As crustal evolution, as opposed to surface evolution, will be influenced most by the larger impact events, only the effects of events forming structures with D > 100 km, have been initially considered. The cumulative number, N_c, of structures, D > 100 km, formed on the earth at 4.6-3.9 b.y. is estimated at 2-3000, from the crater density in the lunar highlands scaled to terrestrial impact conditions and the Phanerozoic cratering rate of the earth scaled to the 4.6-3.9 b.y. flux rate of the moon. For $N_c \propto D^{-2}$(4), the largest structure would be D = 5-6000 km and the cumulative energy, derived from $D = 1.96 \times 10^{-4} (E)^{1/3.4}$ (8), would be $6.5 - 15 \times 10^{28}$J, with the bulk (∿60%) arising from the largest event. Although the average annual rate of energy deposition was only 10-20% of the present annual heat flow on the earth (∿10^{21}J), the energy was deposited as discrete events with each impact D > 100 km releasing virtually instantaneously at a specific location orders of magnitude more energy than the present world-wide annual output.

The most obvious initial effect of the transference of projectile energy to the earth will be to disrupt the proto-crust. Structures with D > 100 km will have final complex multi-ring forms with apparent depths of 2-3 km (9). However, initial depths of excavation were considerably greater, with the amount of stratigraphic uplift in terrestrial craters indicating original depths from which material will be uplifted as > 15 - 20 km for D > 100 km (9). That is lower crustal or mantle materials will be brought to or close to surface, creating lateral compositional and density inhomogeneities in the proto-crust. The uplift of mantle material will favour partial melting at shallow depths and, depending on D, geothermal gradient and composition, may give rise directly to magmatism at the surface. Melting will also be favoured by the addition of post-shock temperatures to the geothermal temperature, ∿ 200°C for pressures of 25-30 GPa at the base of the excavated cavity (10). In addition to promoting direct melting by uplift, impact-induced volcanism is possible through extensive fracturing which may tap partially melted material at depth, with shock wave attenuation data (11) and gravity models (12) indicating fracturing to > 35 km below basins

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with D > 100 km.

The impact basin will also be the site of sedimentation, with possible back-wash of ejecta deposits and rapid degradation of the fractured rim rocks. The model developed so far has points in common with the history of the Sudbury structure, where the formation of a 140km structure at 1.8 b.y. in an area with a high geothermal gradient gave rise to impact-induced magmatic activity (13). However, the thin lithosphere of the early earth, unlike the moon, will not be able to support large mass excesses and the loading of the basin by volcano-clastics and crustal loading by uplift of denser rocks will cause the basin floor to founder. Subsidence of the basin floor down the (still disturbed) geothermal gradient will lead to tectonism, metamorphism and recycling of the inter-basin materials with the production of more fractionated volcanic or plutonic products.

The overall effect of impact is therefore considered to accelerate and localise endogenic activity leading to concentrations of differentiated products from the proto-crust and mantle. The effects of the largest basins will be the most severe and if, as suggested for the moon (14), these major impacts occurred with $\sim$0.1 b.y. the following scenario can be developed. During the period 4.6-3.9 b.y. the earth produced a relatively thin proto-crust which was undergoing widespread but dispersed modification and differentiation by internal processes, promoted by a high flux of impacting bodies up to 10 km in size. At $\sim$4.0 b.y. the earth was subjected to a short-lived series of major impacts by a (separate?) population of bodies up to 10^2km. The resulting impact basins disrupted entirely the local lithosphere and long-lasting ($\sim 10^8$y.) thermal and geologic anomalies were produced. The size and number of basins, $N_c \sim 25$ for $D > 10^3$ km, was such that some overlapped in space and time, resulting in a massive localisation of accelerated mantle and crustal differentiation. This concentration of relatively long-lasting exogenic-triggered endogenic activity over 20-30% of the earth's surface led to the growth of sizeable continental crustal nuclei by 3.8 b.y., corresponding to the first of the accretional "super-events" of the continental crust (15). If this concentration process by major impacts had not occurred, the growth of continental nuclei by dispersed linear zones of high internal activity would not have taken place till later in the earth's history.

This model has similarities to that of (6) and does not favour the creation of proto-oceans by impact (5). The latter is an attractive analogy with the evolution of the moon but results in 40% of the earth's surface at 3.9 b.y. consisting of low density proto-continental crust, which must be completely reworked in a relatively short period by some process other than subduction to agree with the isotopic evidence (15). The model here has the further advantage of explaining why there is no record of the high flux of impacts on earth, the major basins self destruct by endogenic activity, and the absence of rocks older than $\sim$3.8 b.y., a period during which it was not possible to establish sizeable, relatively stable continental nucleii because of the lack of the concentrating effects of the major impacts. The model was developed without establishing a specific proto-crustal composition but would favour a mafic proto-crust as suggested by geochemical evidence (16,17).

Little attention has been paid to the generation of impact lithologies. They may be viewed as mixed lithologies (18) and depending of the size of the event may have compositions intermediate to that of the proto-crust and mantle. They are regarded as surface or near surface rather than crustal lithologies, which will be rapidly modified and destroyed by endogenic processes. Although the bulk of the total volume of impact melt formed on earth will be concentrated in the largest structures, data on the distribution

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of melt at large preserved terrestrial structures (18) suggest that even in the largest structure the melt layer would be < 5 km thick within the final basin. A final point, is that although the upper mantle of the earth is depleted in siderophile elements, it has been suggested that their abundances are in excess of what may be expected from an equilibrium core-mantle separation (19,20). If core separation took place early in the earth's history, prior to the major bombardment, it is possible that the excess siderophiles were added from the impacting bodies at $\sim$4.0 b.y. It has been further suggested that the Archean mantle was enriched in siderophiles over the present mantle (16). Inasmuch as the proto-continents may be the locus of the sites of the largest impact events, it is not unreasonable to speculate that this apparent siderophile enrichment could result from projectile contamination, as observed in some impact melts (21).

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GEOPHYSICAL CHARACTERISTICS OF THE LOWER CRUST IN A PORTION OF THE SUPERIOR PROVINCE IN MANITOBA AND ONTARIO, CANADA. D. H. Hall, Professor of Geophysics and Head, Department of Earth Sciences, University of Manitoba, Winnipeg, Manitoba R3T 2N2.

Recently completed interpretations of crustal structure from seismic, gravity, magnetic and geological data in the Superior geological province in northwestern Ontario and Manitoba show differing styles of crustal structure among several of the subprovinces and major rock units in the area. The data are such as to provide constraints on proposed models for the evolution of the crust. In the present paper, results for the lower crust and the uppermost mantle will be stressed.

1. A seismic reflection survey beneath a large batholith (the Aulneau Batholith) and an adjacent greenstone area reveals a high velocity and presumably high density section lying immediately beneath a strong velocity contrast. This discontinuity, which has been mapped over a wide area by refraction methods, is probably the equivalent of the "Conrad" discontinuity. The high velocity layer varies from 2 to 5 kilometers in thickness over the area mapped, marking the top of the lower crust.

2. The reflection surveys indicate heterogeneous layering in the lower crust with the upper crust much more homogeneous beneath both the batholith and the greenstones. Seismic velocities increase with depth through the whole crustal section.

3. These surveys indicate that the Moho is a layered zone about 8 kilometers thick.

4. Wide angle reflection and refraction surveys have mapped discontinuities in the mantle down to 50 kilometers. One of these discontinuities can be interpreted as a mineral stability boundary if a peridotite composition is assumed. The possible explanation of this and other discontinuities on the basis of other petrological models might aid in discriminating among models.

5. Each of the subprovinces and major rock units tested in the area has a characteristic style of crustal magnetization. In those cases where long-wavelength anomalies suitable for quantitative interpretation occur it is found that the distribution of magnetic sources is usually relatively homogeneous in the upper crust with an increase in concentration towards the bottom of the upper crust or in the uppermost layer of the lower crust. There is some evidence (but not entirely conclusive) that magnetization is widely spread through the lower crustal layer. The study of crustal magnetization is important because the physical properties involved are sensitive indicators of petrological history.

6. Assessment of the capabilities of MAGSAT suggests that the differences in magnetization which frequently are found among subprovinces in the shield can in some cases be resolved by magnetic mapping from satellites. Previous experience has shown that frequency-filtered aeromagnetic data are also suitable for these purposes. One interesting relationship that might be explored in this manner is the correlation which is found between crustal thickness and magnetic field levels on the long-wavelength scale.

7. A decrease in rigidity is evident at 100 kilometers depth beneath the Canadian shield. Relics of the early history of the crust cannot be expected to be found below this level if it is the level at which continental displacement has occurred but may be preserved above it. This upper 100 kilometers of the mantle is a prime target for exploration for such relics, and attempts should be made to demonstrate whether such exploration can be carried out by reflection seismology.

CONTRASTING PETROGENESIS OF GRANITIC ROCKS IN ARCHEAN GREENSTONE AND GNEISS TERRANES, MINNESOTA

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The sources of the granitic intrusions in a 2700 m.y. old greenstone belt-intrusive granite terrane in northeastern Minnesota are compared to those comprising and intruding the 3600 m.y. old Archean gneisses of the Minnesota River Valley, southwestern Minnesota.

Tonalitic and quartz monzonitic rocks of essentially the same age in the greenstone belt have similar low initial strontium isotope ratios but distinctly different trace element patterns. Arth and Hanson, 1975, suggested that the tonalites were derived by partial melting of a basaltic parent at mantle depths, whereas the quartz monzonites were derived by partial melting of graywacke at crustal levels. In both cases the parents would have short histories since the time that their components were derived from the mantle.

The granitic rocks from the gneiss terrane vary from tonalites to quartz monzonites, and include intrusions as young as 1800 m.y. The main granitic phases of the gneisses as well as the later granitic intrusions are leucocratic and characterized by similar subparallel, light REE enriched and heavy REE depleted patterns. It is suggested that the granitic gneisses and later granitic intrusions were derived by partial melting at crustal levels of similar mafic, quartz diorite parents leaving a residue of plagioclase, clinopyroxene or amphibole and garnet.

For these two situations it would appear that for the intrusive granite-greenstone belt in northeastern Minnesota the mantle is the immediate, although indirect, source for the granitic rocks. Whereas for the Minnesota River Valley gneiss terrane the granitic intrusions from 3600 to 1800 m.y. ago were derived from the lower continental crust.

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A COMPARISON OF LUNAR AND TERRESTRIAL ANORTHOSITES. Larry A. Haskin, Karl E. Seifert*, and Marilyn M. Lindstrom, Dept. of Earth & Planetary Sciences and McDonnell Center for Space Sciences, Washington University, St. Louis, MO 63130.

The LPI Planetary Basaltic Volcanism Project has taught us how chemical studies of terrestrial and lunar materials complement each other to improve our knowledge of volcanic processes on both Earth and Moon. Here we suggest how lunar and terrestrial crustal studies might similarly reinforce each other. We compare data for lunar anorthosites and associated rocks with new data on Marcy anorthosites and Tupper-Saranac mangerites, in light of a still tentative interpretation of chemical relationships between those terrestrial materials.

The lunar crust is regarded as ancient and very rich in plagioclase (1,2). This suggestion was supported by the ubiquity of massive anorthosite at the Apollo 16 site. This material is mainly coarse-grained plagioclase (except where shattered by impacts) much like cumulate and adcumulate plagioclase in terrestrial anorthosites (3). Probably, the lunar liquid that crystallized the plagioclase simultaneously precipitated other materials. If so, a physical separation of plagioclase from other minerals occurred, presumably because the plagioclase floated. Similar mechanisms have been proposed to account for terrestrial anorthosite bodies (3).

A lunar crust consisting largely of anorthosite requires plagioclase separation on a grand scale; this led to the hypothesis of a Moon-wide magma ocean (1). As this ocean cooled and plagioclase floated, mafic materials in great volume presumably sank. A few sizeable specimens of cumulate rocks (dunite, norite, troclolite) are present in the lunar sample collection, but not from the Apollo 16 site, which is the only site that may represent typical highlands crust. Although pertinent to the overall picture, these cumulates are rare, are probably not cumulate complements to the anorthosites, and are not further discussed.

Do lunar anorthosites have compositional characteristics expected for products of crystallization of a magma ocean? Can we determine from them the composition of their parent? Important to these questions is material other than plagioclase included within the anorthosites. These materials might represent simultaneously crystallized mafic minerals. However, the grains are small and more likely derived from trapped liquid. Cumulate minerals in terrestrial intrusions invariably trap some parent liquid in its interstices (4). On freezing, this trapped liquid would add to the plagioclase already present plus crystallize mafic and accessory minerals. The chemical behavior of such a liquid during crystallization is known in some detail (5,6) and it is sometimes possible to estimate its composition. Trapped liquid may be parent liquid or more evolved residual liquid, e.g., that on which the plagioclase last floated.

If these rocks were simple mixtures of plagioclase and liquid from a widespread, homogeneous source, their compositions should plot as mixtures of two end members, a straight line on a variation diagram. Some plots (not shown, e.g., FeO-Sc) for a wide variety of highland rocks yield such lines. Also REE distributions show an orderly transition for Apollo 16 rocks, over a range of just a few percent mafic material to KREEP basalt composition. A variation diagram (Fig.1) between two well-determined elements, Sm and Sc, both strongly excluded from plagioclase and presumably present in the trapped liquid component, shows clearly, however, that the highlands samples are not a simple mixture. KREEP basalt cannot be the

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direct parent of anorthosite (7); Fig. 1 shows that it is not a trapped, late-stage supporting liquid, either. It is probably a product of volcanism into highland valleys, there mixed by meteoroid impacts with anorthositic crust. Only rocks of the proposed ANT suite (8) might represent genetic relatives of the anorthosite. At the Apollo 16 site, where anorthosites are abundant, other ANT rocks are mainly breccias, so their initial petrological nature and relationship to the anorthosites has not been determined. Some could be mixtures of plagiocalse and trapped liquid.

The most restricted set of samples that might represent monomict breccias of ANT rocks are those free of meteoritic contamination (9). These are shown in Fig. 1 as triangles; they clearly do not define a line. Does the notion of a single magma ocean need revisions? Or do we now know what chemical characteristics the anorthosite from such an ocean should display? It is naive to presume that any bit of mafic mineral adequately represents average trapped liquid. As trapped liquid crystallizes, part will migrate, leaving early formed crystals in one place, and products of the more evolved liquid in another. This behavior occurs during basalt solidification (10). How much scatter would this cause in a variation diagram? Can we relate such samples to each other?

An approach to these questions is to consider Adirondack-type anorthosite bodies. These are mostly Precambrian, but not as ancient or widespread as lunar anorthosite is thought to be. Many consist of cores of coarse-grained, cumulate plagioclase grading outward into more mafic material. Many are associated with mafic to intermediate rocks (e.g., mangerites) whose genetic relationship to them is debated (3). Chemical evidence has usually seemed not to support any relationship (e.g., 11).

We have analyzed a suite of Marcy anorthosites and adjacent Tupper-Saranac mangerites and tentatively interpret the results as indicating the following relationship: The anorthosite is mainly cumulate plagioclase and frozen trapped liquid. The liquid component may, but does not necessarily, represent the parent from which the plagioclase crystallized. Portions moved during crystallization so that individual analyzed samples do not contain its average composition. The plagioclase and frozen liquid, on a short distance scale, are in metamorphic equilibrium. The mangerite is a differentiated body, with SiO_2 content increasing with height. The estimated average composition of the interstitial material in the anorthosite is the same as that of the mangerite immediately above the anorthosite. The mangerite liquid carried the plagioclase to its present relative position, then was filter-pressed away. It left behind as interstitial material a combination of liquid and crystals of the composition being solidified at the time it separated.

Figure 2 shows the behavior of Sm relative to that of Sc. The anorthosite samples roughly define a mixing line of positive slope, as expected for a mixture of plagioclase and trapped liquid. The mangerite data define a line of negative slope, the lower end representing SiO_2-rich materials that retained Sm but lost Sc to the earlier crystallized minerals (concentrated in the mangerites represented at the higher end). Rough compositions for the trapped liquid component of the anorthosites estimated from partitioning theory approach the composition of the most mafic mangerites, suggesting that they are the same material. There is much scatter, attributable to problems of sampling and use of partitioning theory, but the behavior of all major and trace elements tests and their ratios is consistent with this hypothesis (Seifert et al.in prep. & 12). Isotopic data are also consistent with it (13).

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We are not yet prepared to interpret the lunar data in light of these results. The above shows one way of approaching the lunar data. At this time, too few single specimens of lunar rock have been analyzed for the needed variety of major and trace elements to provide definition of compositional patterns and fluctuations about them.

Acknowledgement: We gratefully acknowledge the partial support of this work under NASA Grant NSG-9055.

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Figure 1

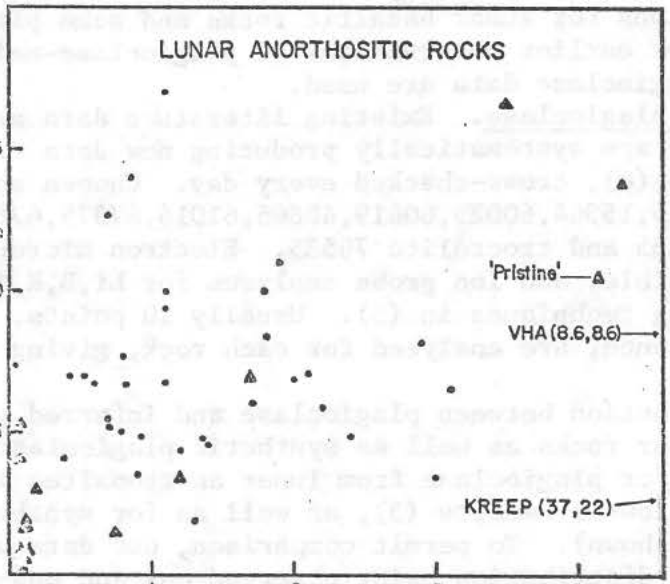
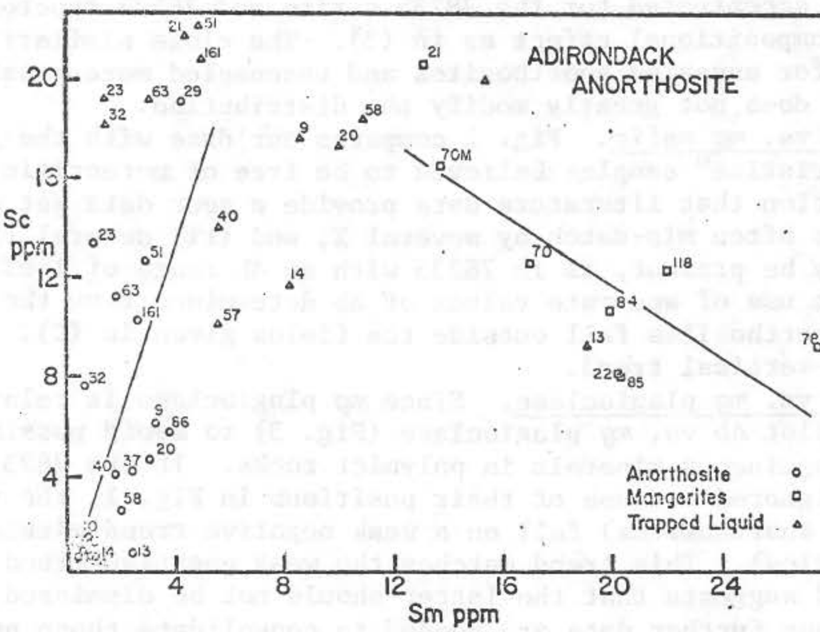


Figure 2



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MINOR ELEMENTS IN PLAGIOCLASE AND MAFIC MINERALS FROM LUNAR PLAGIOCLASE-RICH ROCKS; E.C. Hansen, I.M. Steele, J.V. Smith, Dept. of the Geophysical Sciences, University of Chicago, Chicago, IL 60637.

Plagioclase-rich rocks provide important unresolved clues to the early formation of the lunar crust, but fitting anorthosites into petrologic schemes is difficult because a) anorthosites have higher Fe/Mg than other primitive rocks, b) a plot of Ab(plagioclase vs. *mg*=atomic 100Mg/(Mg+Fe) of coexisting mafic minerals is not uni-valued, and c) the processes for accumulation of anorthosite bodies are controversial. Particularly troublesome is distinguishing between surviving homogeneous rocks and relics of mixing, melting and metamorphism, though low abundance of siderophile elements (1,2) may identify the former.

We provide new minor-element analyses of plagioclase, mainly from plagioclase-rich rocks, to a) obtain the distribution of Fe/Mg between plagioclase and mafic minerals in plutonic and annealed samples, b) compare it with distributions for lunar basaltic rocks and some plutonic rocks (3), and c) test whether earlier observations of plagioclase-mafic distributions hold when only plagioclase data are used.

Minor elements in plagioclase. Existing literature data are mostly unreliable, and we are systematically producing new data referenced to reliable standards (4), cross-checked every day. Chosen samples are mainly anorthosites (15415,15364,60025,60619,60665,61016,67075,67637,67746) but include norite 78235 and troctolite 76535. Electron microprobe techniques are used whenever possible, and ion probe analyses for Li,B,K,Ti,Cr,Mn,Ba and Sr are scheduled using techniques in (5). Usually 10 points, chosen to avoid secondary fluorescence, are analyzed for each rock, giving the mean and 1σ range in Table 1.

The *mg* distribution between plagioclase and inferred parent liquid was consistent for lunar rocks as well as synthetic plagioclase-liquid pairs (3). Our data (Fig. 1) for plagioclase from lunar anorthosites follow a similar trend to that for low-Ti basalts (3), as well as for synthetic plagioclase-liquid pairs (not shown). To permit comparison, our data were converted to a plagioclase-liquid distribution using observed opx and opx-ol (7) and ol-liquid (11) distributions. The slight displacement for the anorthositic plagioclase is accentuated for the 78235 norite and 76535 troctolite, perhaps indicating a compositional effect as in (3). The close similarity of distributions for annealed anorthosites and unannealed mare basalts suggests that annealing does not greatly modify the distribution.

Ab plagioclase vs. *mg* mafic. Fig. 2 compares our data with the generalized regions for "pristine" samples believed to be free of meteoritic contamination (1,2). We caution that literature data provide a poor data set since (i) Ab and An contents often mis-match by several %, and (ii) several types of plagioclase may be present, as in 78235 with an Ab range of 1.6% (6). We urge consistent use of accurate values of Ab determined from the Na_2O content. Three of our anorthosites fall outside the fields given in (1). All our data fall in a near-vertical trend.

Ab plagioclase vs. *mg* plagioclase. Since *mg* plagioclase is related to *mg* mafic, we now plot *Ab* vs. *mg* plagioclase (Fig. 3) to avoid possible problems from non-consanguineous minerals in polymict rocks. If the 78235 and 76535 specimens are ignored because of their positions in Fig. 1, the remaining specimens (all anorthosites) fall on a weak negative trend within their 1σ errors (rectangles). This trend matches the weak negative trend in Fig. 2 and in (2), and suggests that the latter should not be dismissed as in (1).

Conclusion. Many further data are needed to consolidate these preliminary

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Table 1. Minor elements in plagioclase: *mg* values.

Sample	Ab mol %	FeO wt %	MgO wt %	K_2O wt %	*mg* plag	*mg* ol	*mg* px
15415,89	3.5(1)	.105(11)	.047(6)	.025(5)	44(3)	-	[a]56
15364,1	5.2(8)	.058(5)	.046(6)	.040(6)	59(2)	71	76
60025,108	3.9(5)	.14(3)	.064(12)	.031(11)	44(4)	-	55
60619,12	5.5(13)	.11(3)	.10(3)	.03(1)	61(4)	72	75
60665,3	3.6(4)	.13(4)	.06(3)	.016(6)	42(4)	67	60
61016,15	3.5(3)	.11(2)	.07(1)	.006(3)	54(4)	-	62
61016,27*	3.8(4)	.231(10)	.17(7)	.022(12)	56(4)	-	"
61016,27+	3.8(6)	.121(12)	.051(10)	.023(6)	42(4)	-	"
67075,2	3.1(3)	.069(18)	.042(15)	.020(7)	52(7)	[b]60	-
67637,1	5.4(4)	.100(17)	.048(15)	.014(3)	45(6)	60	64
67746,1	5.7(8)	.071(9)	.088(18)	.084(17)	69(3)	75	79
76535,49	3.3(3)	.045(12)	.19(4)	.051(4)	87(3)	[c]87	[c]85
78235,40	4.7(2)	.021(2)	.059(11)	.071(18)	77(9)	-	[c]78

Standard deviation in brackets. *and +, melted and non-melted portions.
a,b,c, respectively from refs. 8,9 and 10.

results, especially ion-probe data for additional elements. The consistency of the *mg* partitioning suggests that post-crystallization processes have not eliminated clues to the origin of lunar anorthosites.

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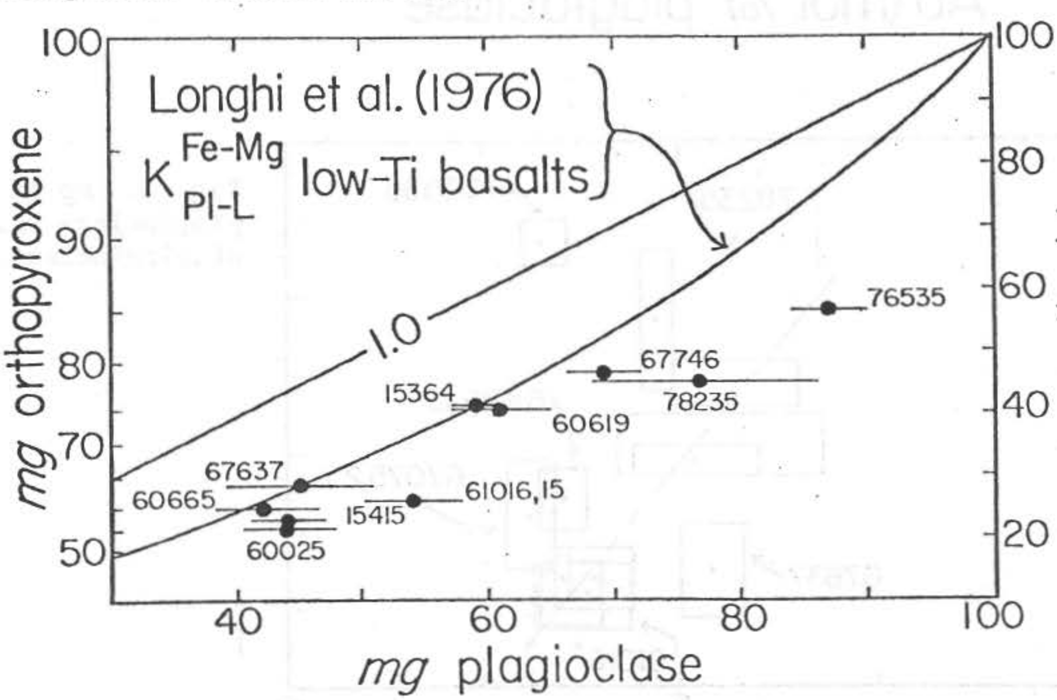

Fig. 1. Distribution of Fe and Mg between low-Ca pyroxene and plagioclase and comparison with plagioclase-liquid data of (3).

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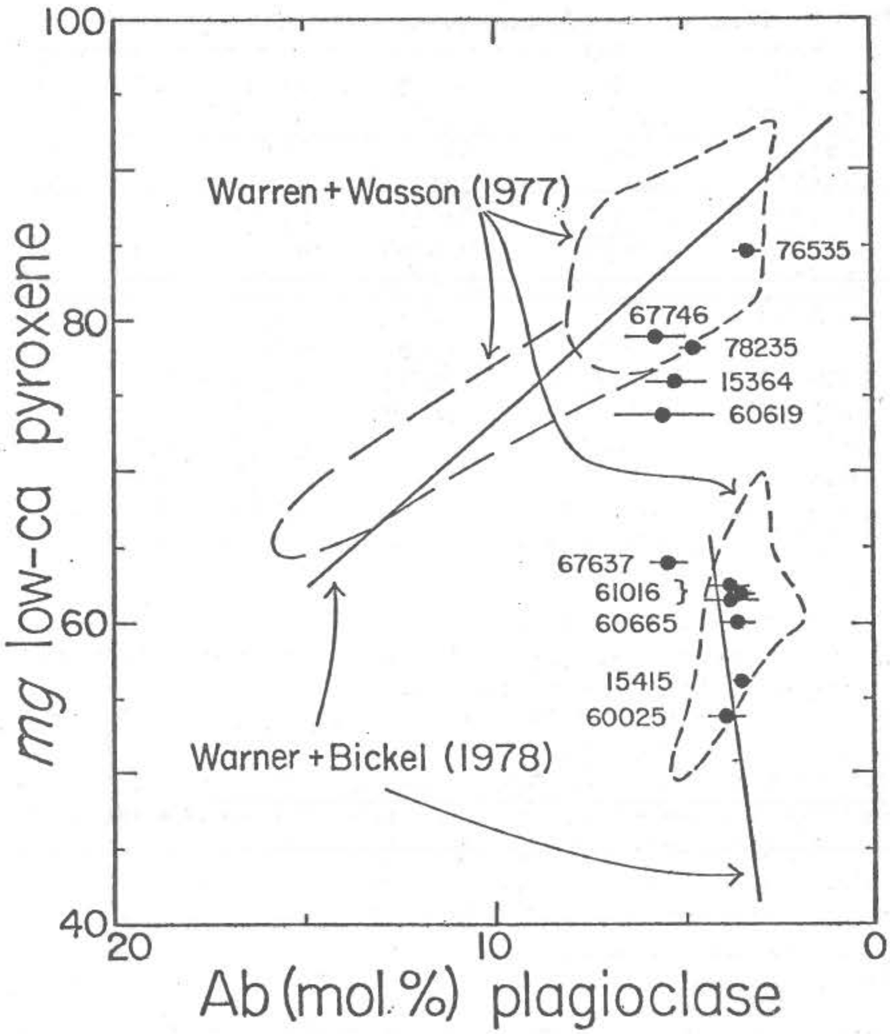

Fig. 2. Plot of *mg* (low-Ca pyroxene) vs. *Ab* (mol %) in plagioclase. Our data use mg for low-Ca pyroxene, but the data in Warren and Wasson (1977) are for both olivine and pyroxene. The dashed lines are our interpretation of their data. The trends from Warner and Bickel (1978) pass through data for both pyroxene and olivine.

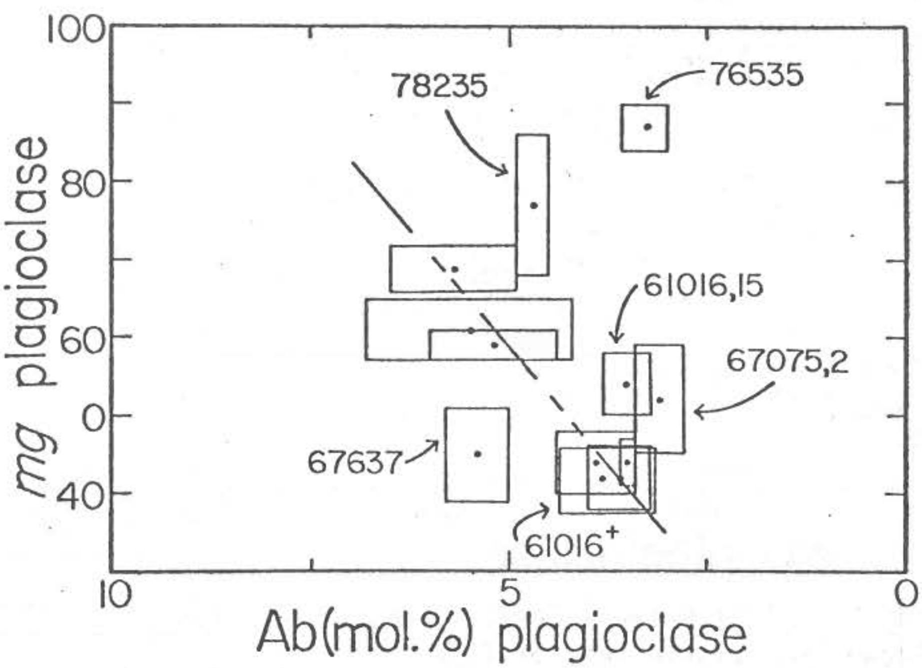

Fig. 3. *mg* of plagioclase vs. *Ab* of plagioclase.

XENOLITH STUDIES AND THE NATURE OF THE LOWER CRUST, R. W. Kay, 304 Kimball Hall, Cornell University, Ithaca, NY 14853

Xenolith populations in basalts and kimberlites include numerous examples of both upper mantle and lower crustal rocks. Of the two types, the mantle suites have received more attention, and serve as a guide to the study of the crustal suites. In the lower crust, there are several rock types that we expect to be important, and are found among the xenoliths:

(1) Residual granulites complimentary to crustal melts.
(2) Intrusive basaltic magmas, which serve as heat sources.
(3) Metasedimentary rocks.

Many of the xenoliths are pre-Cambrian, but largely by chance.

Present heat flow variation in the U. S. implies that lower crustal rocks will be at widely different present-day temperatures. Temperatures exceed the melting temperature of crustal rocks for many areas in the Western U. S. Xenoliths from these hot areas (e.g., Rio Grande rift) have P-T estimates that agree with the present day temperature profile. In cooler regions, the mineral equilibration temperatures probably do not refer to present-day conditions, but to fossil geothermal gradients. Retrograde mineralogies, as in the xenoliths from the Colorado plateau, are superimposed on the high temperature relict mineralogy.

Sampling problems plague xenolith studies. One of the most significant of these is the question of how to regionalize observations from a limited number of xenolith sites. Geophysical techniques can be useful in extrapolation to a regional scale. Among these techniques, travel time residuals, magnetic and conductivity data, and seismic reflection profiling can be used in conjunction with xenolith data to develop regional models of the lower crust. Some apparent contradictions remain to be resolved. For example, the high electrical conductivity of some pre-Cambrian lower crustal regions seems to be incompatible with commonly inferred anhydrous mineral assemblages.

Lower crustal xenoliths from presently active volcanic arcs (e.g., Japan) give the clear implication that basaltic rocks are the major rock type that is being added to the lower crust at the present. In fact, basic igneous rocks seem to be quite common in all lower crustal xenolith populations. The relevance of models of magma formation at volcanic arcs to the question of the mechanisms of continental growth depends on the present and past rates of crustal formation and recycling.

PLATE TECTONICS ON THE EARLY EARTH; W.S.F. Kidd and Kevin Burke,
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Heat escapes from terrestrial planets both by conduction through the surface and by convection. On earth's surface the plate structure of the lithosphere with its associated volcanism and the occurrence of volcanoes unrelated to plate structure are the most obvious indications of convection within. On the moon although convection may persist today, the absence of any volcanism younger than about 3 b.y. means that there is no superficial sign of young convection. On this evidence the moon could have completed its convectional evolution mainly during the time of high impact flux (> 4 b.y.). This process would have needed to concentrate the moon's heat generating nuclides close to the surface so that conduction could suffice to remove heat generated later in lunar history.

On Mars there is evidence of both early and later vulcanism (although how late vulcanism persisted is not yet clear) so that convection, on the basis of superficial evidence, persisted for longer than on the moon. Because there is only local evidence of compressional and strike-slip motion on Mars the strong evidence of rifting can be interpreted as implying that the planet came close to, but did not develop, a plate-structured lithosphere. Mars appears to have been generally able to dissipate its convective heat through sporadic vulcanism.

On Earth (1) about one-third of the heat presently being generated escapes through the cooling of ocean floor as it ages, another third by conduction through the ocean floor and the rest by conduction through the continents in which there is a high concentration of heat-generating nuclides at shallow depths. Evidence of high sea levels during the last 0.6 b.y. has been interpreted (1) as showing that at some times in the past up to half of all the earth's heat production was lost by the aging of ocean floor.

During the Archean two or three times as much heat was being generated in the earth and the absence of evidence of extensive continental melting (2) shows that heat did not escape along much steeper conductive gradients than at present. Evidence of incessant widespread sporadic vulcanism is also lacking so that increased convection with more effective cooling of aging lithosphere appears the most likely way by which the extra heat was removed.

Plate tectonics is the heat dissipating convective process that operates on earth today, and we have suggested (3) that Archean rocks and structures can be interpreted as products of plate tectonic processes. More active plate tectonics could have satisfied the heat dissipating requirements of the early Earth.

Some Archean geologists have been reluctant to interpret the rocks they study as products of plate tectonic processes and some alleged plate tectonic interpretations of the Archean appear unrealistic to us. This seems to be partly through a lack of familiarity among students of the early Earth with the way in which plate tectonics has operated during later times and limited acquaintance with the kinds of rocks and structures it has produced. A major specific problem has been that many Archean geologists interpret greenstone-belts as structurally simple whereas plate tectonic interpretations, emphasizing horizontal motion on the Earth's surface, require them to be structurally complex as Ramsay (4) long ago demonstrated for the Barberton mountain land.

The plate tectonic process on Earth embodies, as an essential part, the convective return of hydrated lithosphere to the mantle. Partial melting at or above subducted hydrated lithosphere is the process that produces the calc-

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alkaline rocks characteristic of convergent plate boundaries and many have pointed out that lateral accretion of such arc material made at convergent boundaries is a feasible way of building continents. These observations suggest a possible answer to the question: "When did plate-tectonics begin?" One answer could be "As soon as water, sufficient to partially hydrate descending convective material, existed on the Earth's surface, because from that time on calc-alkaline rocks are likely to have been erupted." We suspect that water existed on the Earth's surface from early in its history - certainly long before the ~3.8 b.y. oldest rocks were formed. From the time that there was water on the Earth's surface there seems little likelihood that terrestrial petrology much resembled that of the moon. Instead of high alumina basaltic and anorthositic highlands the Earth would have produced calc-alkaline arcs, microcontinents and continents.

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THE THERMAL HISTORY OF THE EARTH IN THE ARCHEAN; R. St J. Lambert, Department of Geology, University of Alberta, Edmonton, Alberta T6G 2E3

It is argued that the Earth is in thermal equilibrium, with heat production roughly equal to surface heat flow. To produce an average flux of 60 mW/m^2 a mantle with about 245 ppm K, 0.028 ppm U and Th at 0.080 ppm is required if there is zero heat production in the core. Such a mantle could comfortably produce tholeiites of the type found in Leg 37 of DSDP, drilling into the flank of the Mid-Atlantic Ridge. The thermal equilibrium argument can be extended back to the Archean-Proterozoic boundary, but no further, for thick Archean continents could only stabilize if the mantle-crust heat flux is limited to about twice today's value. The paradox of lack of extensive pre-3000 Ma continents may be resolved if a model with a heat flow maximum is adopted; the double convection systems of McKenzie and Weiss come close to providing such a solution. Arguing further that high heat flows today are invariably associated with igneous activity, it is reasoned that all the Archean continents could be produced in only 60 Ma if the surface heat flow ever averaged 180 mW/m^2, three times today's average or about equal to Iceland. Concluding that such an event has never occurred, together with other analogies, leads to a heat flux model in which a maximum of 125 mW/m^2 is reached at 2800 Ma ago, preceded by a subsidiary maximum at 3500 Ma. Rapid continental growth occurred from 3000 to 2600 Ma, leading to stabilization as the heat-producing elements were transferred from mantle to crust. After consolidation of the scattered sial, the sub-continental mantle began to develop a lithosphere, leading to a period of "thin-skin tectonics" (the Proterozoic). Finally, when some of the oceanic lithosphere reached a thickness comparable with that of today, the plate tectonic regime began. The only alternative seen by this author to the above is to recycle the Archean proto-continents extensively, a process for which we have no direct evidence.

THE ORIGIN OF LUNAR ANORTHOSITES

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Lunar anorthosites are a distinct class of rocks, possessing a combination of unusual petrological and chemical features that suggest formation of the anorthosites under rather special conditions. Among the ancient, lunar crustal rocks which retain textural evidence of a cumulate origin and which are free of meteorite contamination, the anorthosites form a distinct group in terms of the compositions of plagioclase and the mafic minerals(1,2). In the suite of Mg-rich cumulate rocks (dunite,norite,troctolite) there is a conventional trend of decreasing anorthite content in plagioclase (An 98-86) coupled with decreasing Mg/(Mg+Fe) in olivine and low-Ca pyroxene (Fig. 1). Among the anorthosites, however, there is less than 3 mole % variation in plagioclase composition (An 98-95), while mafics vary approximately 25 mole % in Mg/(Mg+Fe) (Fig. 1). In most anorthosites the variation of Mg/(Mg+Fe) in olivine and pyroxene is no more than a few mole percent, but some samples, eg. 67075 (3) and 60025(4) show much larger ranges (25 and 7 mole % respectively) indicating a possible mixing process. There is a distinct gap in mineral compositions between the anorthosites and the Mg-rich cumulates, which was originally thought to be an artifact of limited sampling (5), however, the gap has persisted despite the addition of many new samples(4) to the original groupings of (1). Thus the gap is apparently real and must be explained. Furthermore, even though the anorthosites have mafics with much lower Mg/(Mg+Fe) than Mg-rich cumulates containing plagioclase of similar composition and thus appear to be more differentiated, extremely low Sr^{87}/Sr^{86} ratios in the anorthosites (6) indicate that the anorthosites formed within a few million years of the formation of the moon, whereas the primary differentiation of the moon lasted 100-200 million years.(7). Also, studies of ejecta from basin-forming impacts(8) suggest that the lunar crust is stratified with the upper layers richest in anorthosites and the lower layers rich in Mg-rich cumulates. These data beg the paradoxical interpretation that the anorthosites formed before the Mg-rich rocks.

The general petrological and chronological features of the primitive crustal rocks can be accomodated by the "rockberg" model of the formation of the crust (9,5). This model assumes that the outer portion of the moon was initially molten, i.e. a magma "ocean" existed. Strong radiational cooling induced extensive crystallization of the upper layers. Dense olivine and pyroxene sunk, while plagioclase with density $\lesssim$ density of the fractionated liquid floated and aggregated over cooler areas of convective down-welling to form anorthositic "rockbergs" (Fig. 2). During this episode the magma ocean would crystallize rapidly without much opportunity for fractional crystallization except in its upper layers(10). When the ocean had crystallized sufficiently to be saturated with plagioclase,lateral accretion of the rockbergs into a continuous, thin crust would have come to completion quickly and the downward growth of the crust would have accelerated(11), thereby allowing the incorporation of less extensively fractionated and, hence, more Mg-rich material into the lower crust.

Attempts to model the mineral compositions in the primitive crustal rocks (5,12) indicate that the actual circumstances involving the crystallization of these rocks were quite complex. The trend of mineral compositions produced from calulated fractional crystallization (5,12) of model magma ocean composition GA(V) (solid line in Fig. 3) passes at a marked angle to the anorthosite trend. This line represents the expected mineral compositions if both mafics and plagioclase were cumulate or heteradcumulate phases. If the mafics formed entirely by crystallization of intercumulus liquid, the composition of the

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plagioclase would still be dominated by the cumulus plagioclase, but the Mg/(Mg+Fe) of the mafics would be that of the trapped liquid, since this liquid would be unable to differentiate. The trace of anorthosite mineral compositions produced by liquid entrapment is shown as the dashed line in Fig. 1. Combinations of cumulus mafics and trapped liquid would produce anorthosites with mineral compositions lying between the two curves. The natural case is likely to have been more complex with less efficient fractionation driving the curves toward higher Mg/(Mg+Fe) and efficent separation of mafics coupled with suspension of plagioclase steepening the curves. However, in general we should expect to find evidence of greater proportions of trapped liquid, such as higher concentrations of incompatible elements, with decreasing Mg/(Mg+Fe). This prediction is borne out by the light REE. Fig. 3 shows that bulk rock La and Sm concentrations increase with decreasing Mg/(Mg+Fe) of the mafic minerals.

The limited range of plagioclase composition in the anorthosite series might well be the result of the small density contrast between plagioclase(ρ= 2.76) and the liquid (Fig. 3). Given the turbulent convection in the upper layers of the magma ocean, early formed plagioclase might have remained suspended and hence in equilibrium with the liquid until removal of olivine drove the Fe/Mg ratio and density of the liquid to higher values. Predominantly equilibrium crystallization of plagioclase could account the narrow range of plagioclase composition of the anorthosites and variable proportions of cumulus mafics and trapped liquid could produce the anorthosite "trend". This model is different from that of (13) in that plagioclase suspension is restricted to the upper layers of the magma ocean and the mixing of cumulus mafics and trapped liquid accounts for the anorthosite trend only -- it does not relate the anorthosites to the Mg-rich rocks which crystallized from different liquids.

Relating the anorthosites to the Mg-rich cumulates presents a considerable problem if we assume the same parental liquid, i. e. an initially homogeneous magma ocean. A parental liquid that can produce the anorthosites, GA(V), is too Fe-rich to produce the Mg-rich rocks by simple fractional crystallization and accumulation (Fig 1). Equilibrium crystallization of GA(V) would produce the narrow range of mineral compositions given by the line "eq". If composition GA(V) underwent near equilibrium crystallization up to the point of plagioclase saturation and then underwent fractional crystallization, the trace of mineral compositions would originate at the right-hand end of the line "eq" and continue through the low Mg/(Mg+Fe) portion of the Mg-rich cumulate area. In order to accomodate all of the Mg-rich cumulates it is necessary to assume less efficient fractional crystallization or an initial composition with higher Mg/(Mg+Fe). The presence of cumulus olivine in some of the anorthosites (2), however, suggests that the initial bulk composition cannot have been very much more magnesian if the anorthosites were primary.

The formation of the primitive crustal rocks, then, may be considered to be the result of two episodes in the evolution of the magma ocean: the first involved formation of anorthositic rockbergs in the highly fractionated upper layers of a magma ocean that was undergoing rapid, near-eqilibrium crystallization; the second involved downward growth of the crust into an ocean saturated with plagioclase and undergoing fractional crystallization.

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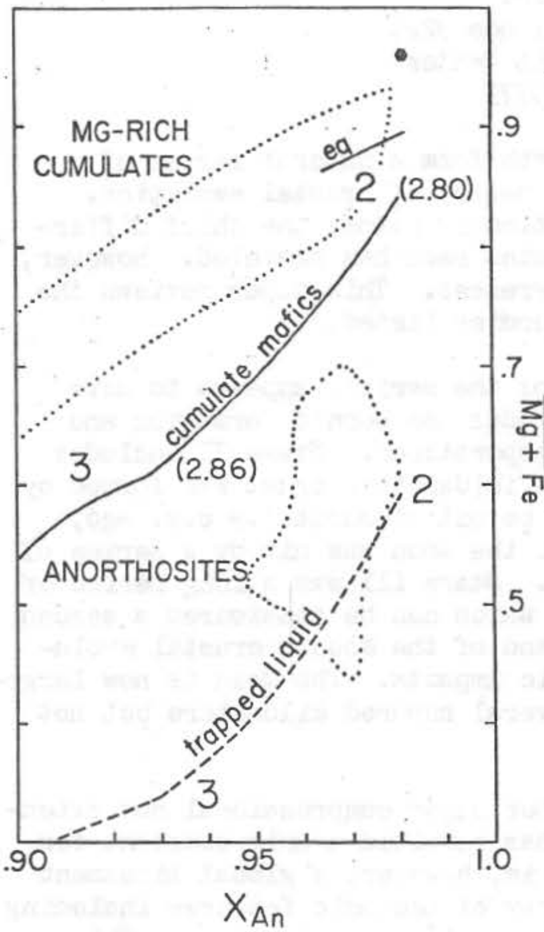


Fig. 1 Mg/(Mg+Fe) in olivine and low-Ca pyroxene versus anorthite content in plagioclase for primitive lunar crustal rocks. Observed compositions lie within the dotted lines. Mineral compositions expected from equilibrium crystallization lie along the line "eq", from fractional crystallization along the line "cumulate mafics" and from cumulus plagioclase with trapped liquid during fractional crystallization along the line "trapped liquid". "2" refers to the onset of plag+ol crystallization; "3" to plag+low-Ca pyroxene. Numbers in parentheses are calculated liquid densities.Model parental composition GA(V) in wt.%: SiO_2 -42.8, TiO_2 -0.44, Al_2O_3 -8.36, Cr_2O_3 -0.17, FeO -9.40, MgO - 33.6, MnO - 0.03, CaO -4.88, K_2O -0.01, Na_2O -0.07.

Fig. 2 Sketch of the "rockberg" epoch of the lunar magma ocean.dots=ol, crosses = opx, x's = cpx, squares = plag. Not to scale though ocean depth estimated to be approx. 300-400 km.

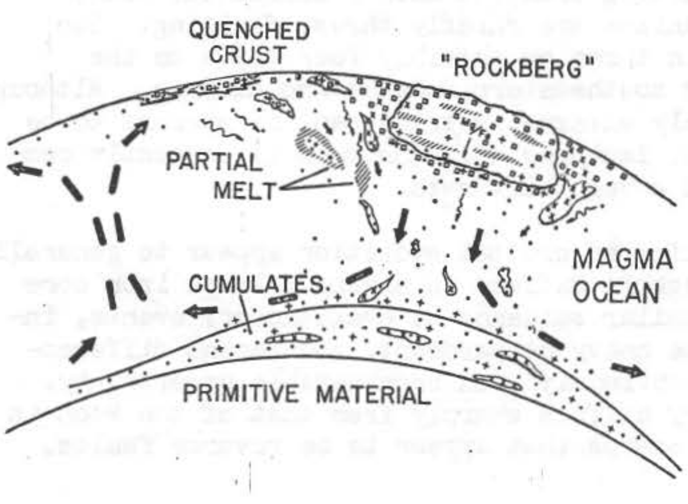
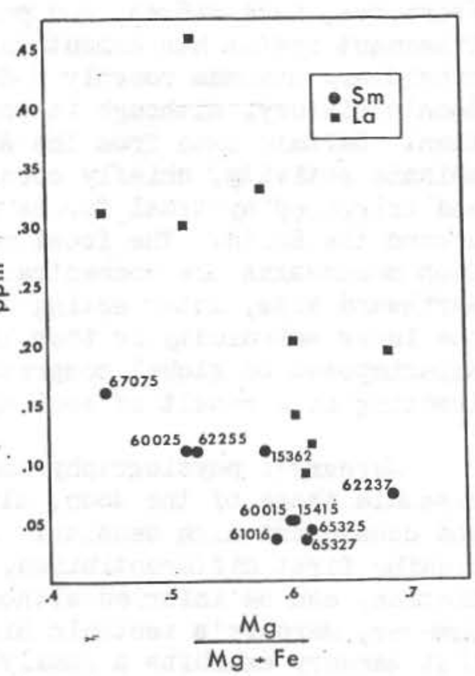


Fig. 3 La and Sm bulk rock concentrations in lunar anorthosites as a function of Mg/(Mg+Fe) in mafic minerals. Where mafics are inhomogeneous, e.g. (3),most Fe-rich compositions are plotted.



A REVIEW OF TECTONIC PROCESSES IN THE INNER SOLAR SYSTEM

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The Moon, Mercury, Mars, Venus, and Earth form a natural series of increasing size, mass, internal energy, and degree of crustal evolution. They appear to have undergone similar evolutionary paths, the chief difference among them being how far along these paths each has traveled. However, their tectonic styles show substantial differences. This paper reviews the tectonic processes dominant on each of the bodies listed.

The Moon, smallest and most primitive of the series, appears to have evolved in three major stages. Stage I includes the Moon's formation and the heating of its outer parts to melting temperatures. Stage II includes the first differentiation, in which a global feldspathic crust was formed by igneous processes. The end of Stage II can be put at about 3.9 b.y. ago, the time of the late heavy bombardment, when the Moon was hit by a series of massive bodies which formed the mare basins. Stage III was a long period of generation and eruption of basaltic magmas, which can be considered a second differentiation. This was essentially the end of the Moon's crustal evolution, except for minor volcanism and sporadic impacts. The Moon is now largely inactive, cold and rigid to depths of several hundred kilometers but hot and partly molten in the deep interior.

The Moon's physiography indicates neither major compressional nor extensional tectonism, suggesting that its size has remained nearly constant for the last four billion years or more. There is, however, a global lineament system, or lunar grid, consisting of a variety of tectonic features including fractures, mare ridges, and possibly volcanic chains or extrusions. This lineament system has azimuth maxima in NW-SE and NE-SW directions, with a subsidiary maximum roughly N-S. It appears to date from very early in the Moon's history, although it apparently has been reactivated locally since then. Seismic data from the ALSEP seismometer net indicate a low level of seismic activity, chiefly occurring at depths of several hundred kilometers and triggered by tidal forces resulting from the Moon's elliptical orbit around the Earth. The focal mechanisms are chiefly thrust faulting. The deep moonquakes are concentrated in three or possibly four belts on the earthward side, intersecting under southeastern Oceanus Procellarum. Although the lunar seismicity is thus largely externally triggered, it appears to be superimposed on global compression, implying that the Moon is presently contracting as a result of cooling at a very slow rate.

Mercury's physiography and inferred crustal evolution appear to generally resemble those of the Moon, although it differs in having a large iron core and consequent high density. A similar sequence of evolutionary events, including first differentiation, late heavy bombardment, and second differentiation, can be inferred although obviously with considerable uncertainty. However, Mercury's tectonic history differs sharply from that of the Moon in that Mercury exhibits a family of scarps that appear to be reverse faults.

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They show no preferred orientation, unlike the lunar fracture pattern. Their age is largely pre-late heavy bombardment (i.e., they are older than the Caloris basin). Two mechanisms have been suggested for the formation of the inferred reverse faults: planetary despinning to the present resonant rotation, and simple thermal contraction. The random orientation at all latitudes suggests that thermal contraction is by far the dominant mechanism. The timing and nature of Mercurian tectonism suggests higher early temperatures than those in the Moon, perhaps resulting from Mercury's position closer to the sun.

The physiography of Mars has been greatly modified by atmosphere-dependent processes of erosion and deposition. However, if allowance is made for these, a grossly similar overall pattern of crustal evolution can be discerned, including first differentiation, late heavy bombardment, and second differentiation. Unlike the Moon and Mercury, Mars continued into a fourth stage of crustal evolution that can be considered incipient plate tectonics. The most conspicuous evidence for this inference is the Valles Marineris, generally interpreted as an immense system of grabens and related features resulting from uplift and N-S extension. The Valles Marineris are superimposed on the N-S trending Thaumasia uplift, a similar but older and less-developed feature. The Valles Marineris appear fundamentally similar, if allowance is made for simpler crustal structure and greater lithospheric thickness, to the East African rift system, and represent in this view the beginning of crustal fragmentation by tectonic processes. However, their great age suggests that Mars has not advanced very far into a true plate tectonic stage. This is supported by the great size of the Olympus Mons volcanic pile, which has been interpreted as resulting from continued eruption on a crust that has remained stationary over a mantle plume.

Only portions of the physiography of Venus have been resolved by earth-based radar. Some regions are cratered, resembling the lunar, Mercurian, and Martian highlands to an extent. However, a large polygonal basin, a plateau, and a long valley with a bifurcation at one end have also been resolved, suggesting that a substantial part of the Venusian physiography is of tectonic origin. The valley is of particular interest, having a gross resemblance to the Red Sea and the East African rift system, even to the extent of a possible triple junction. Geochemical data from three Soviet Venera probes suggest that differentiation has occurred on Venus. Viewed collectively, the data suggest that Venus, like Mars, is a transition planet, which has entered a fourth, or plate tectonic, stage of crustal evolution.

The Earth is obviously the most active and highly-evolved member of the inner solar system, and its early geologic record is obscure for that reason. However, terrestrial Precambrian geology, when viewed in the context of comparative planetology, can be interpreted as resulting from an evolutionary path similar to that followed by the other planets and the Moon. It has been proposed that the Earth underwent early differentiation, forming a global crust of intermediate bulk composition, in the first half-billion years of its existence. A period of late heavy bombardment, similar to that suffered by the Moon, disrupted this early crust, localizing mantle convection and basaltic magma generation, and initiating sea-floor spreading analogous to, but much more rapid than, that of the Phanerozoic. This

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permobile stage probably ended about 2.5 billion years ago, with crustal thickening and stabilization. Since that time, terrestrial tectonism has been dominated by sea-floor spreading, subduction, and attendant processes, although intra-plate activity has played a continuing role whose extent is still not known. Satellite imagery and seismic patterns indicate that the degree of intra-plate deformation and extent of diffuse plate boundaries are much greater than commonly realized. A petrologic effect of the Earth's intense tectonic activity has been redifferentiation of the original crust by partial melting, resulting in a vertically-zoned continental crust consisting of a depleted granulitic lower part and a lithophile-rich granitic upper part. The Earth's continents in this theory represent the greatly-altered remnants of a primordial global crust, rather than the result of long-term lateral accretion of orogenic belts.

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YET ANOTHER ARCHEAN PROVINCE? THE LOWER CRUST AS REVEALED BY XENOLITHS IN KIMBERLITES, T. R. McGetchin, Lunar and Planetary Institute, 3303 NASA Road 1, Houston, TX 77058

Introduction

While our knowledge of the nature of the Archean is through classical field studies of old terrains exposed in the shields, the crust of the earth is also available as xenoliths in kimberlite pipes. These rocks are of particular interest because in some pipes a very long vertical section is represented which includes not only the crystalline rocks known from the upper crust, but also xenoliths from the lower crust and upper mantle. Xenoliths of possible upper mantle origin (eclogite and lherzolite nodules) have been the subject of intensive study and a great deal is known about them. The lower crustal suite has *not* received this type of attention and these rocks may provide very important clues about the nature and origin of the crust of the earth (and whether the processes which formed it had anything in common with the lunar uplands).

Setting and Xenoliths

The kimberlite pipes of the Colorado Plateau occur in southern Utah, Arizona, and New Mexico (see Figs. 1,2). Pyroxene compositions from within the intrusive kimberlite microbreccia suggest depths 50 to 200 km. The kimberlite is known to be derived from two lherzolite assemblages by simple physical dis-aggregation, (a) spinel lherzolite and (b) garnet lherzolite, presumably within the upper mantle. The pipes and dikes are filled with breccia. At Moses Rock, 3% of the breccia consists of crystalline rock fragments, approximately 75% of these rocks are igneous or altered igneous rocks — metabasalt, gabbro, diorite, rhyolite and granite; 25% of the fragments are foliated metamorphic rocks. The most abundant metamorphic type is a coarse-grained, granular-textured garnet-bearing, hornblende, plagioclase (gabbroic) gneiss, probably metagabbro. Detailed field counts of all xenolithic fragments suggest that fragment size is inversely related to depth of origin — apparently rocks are comminuted during vigorous transport up the vent during eruption. On this basis, with supporting petrographic studies, a model for the volumetric abundance of rocks to about 100 km depth has been proposed by McGetchin-Silver (1972) (see Fig. 3). Of particular interest are the granulite gneisses of the lower crust. Recently Padovani *et al.* (1978) have proposed a model (very similar to Fig. 3) based on laboratory measurements of acoustic velocity in these rocks.

The Lower Crust

The lower crust under the central Colorado Plateau (see Fig. 3) is believed to consist of garnet-gabbroic gneiss and possibly eclogite. These rocks show a hydrous over print. In contrast, Padovani and Carter (1977) report an anhydrous assemblage of xenoliths from Kilborne Hole, New Mexico, just off the SE margin of the Plateau within the Rio Grande Rift. They report sillimanite and opx bearing garnet granulite, charnockite, two-px granulite and anorthosite. They infer depths of 22 to 28 km and temperatures near 800°C to 900°C for the garnet-plagioclase bearing rocks. Hence, the assemblages at the two localities may be similar; they are strikingly different in the degree of hydration.

Some Implications

The events recorded in volcanic rocks and structural deformation at the surface clearly leave their imprint on the lower crust where (at least part of) the action goes on. The hydration of the lower crust of the Colorado Plateau is related to the regional Tertiary tectonic and igneous history of the Western U. S. The sequence of events in the central Colorado Plateau in Tertiary time was (a) kimberlite emplacement at about 30 m.y., (2) just subsequent (or possibly contemporaneous) emplacement of minette (K-rich lamprophyre) dikes, (3) laccolith emplacement, generally dioritic in composition, scattered in time, (4) structural uplift, at about 18 m.y., and (5) pervasive bimodal basaltic-rhyolitic volcanism around the margin of the plateau from about 30 to the present. Apparently heat (and possibly volatiles) were released beneath the Colorado Plateau during subduction from the west; the upper mantle released volatiles (kimberlites), partially melted (minettes), expanded due to heating from below (regional structural uplift), and locally partially melted the lower crust (volcanic rocks). Padovani and Carter interpret the anhydrous granulitic lower crustal rocks (from the Southern Rio Grande Rift) to be partial melting residues after removal of felsic lavas. Smith (1977) described hydrated peridotites from Green Knobs and Buell Park (see Fig. 2) and also suggested that the hydration event is related to the gneiss of kimberlite and pervasive mantle hydration. Smith suggests that the dehydration of this mantle by intrusion of minette magma may have liberated the volatiles which caused the violent kimberlite emplacement. There are several generalizations to be drawn:

(1) The lower crust is yet another probable pre-Cambrian (Archean?) terrain available for study (xenolithic rhyolites from Moses Rock have been dated at 1.7 b.y.).

(2) Xenoliths suggest the lower crust is predominantly mafic, metaigneous garnet-gabbroic gneiss which, however, is quite variable over rather short distances in terms of hydration.

(3) The lower crust participates in igneous events, hence is almost certainly pervasively reworked.

Finally, the composition and state of the lower crust of the earth, and processes active there may be important clues about the origin of the crust itself. Age dates on these rocks are badly needed.

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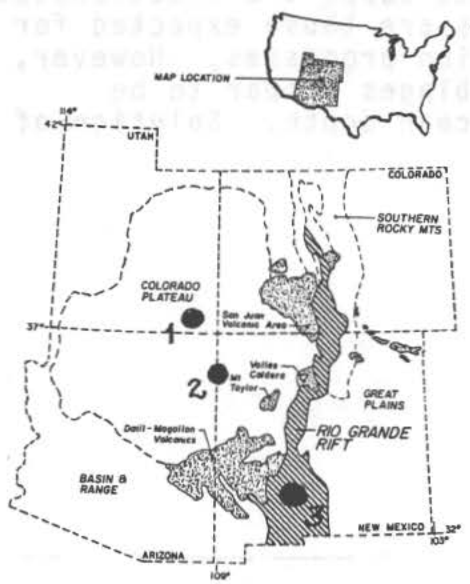
McGetchin, T. R.

1 Moses Rock
2 Buell Park
3 Kilborne Hole



The Rio Grande Rift

Figure 1

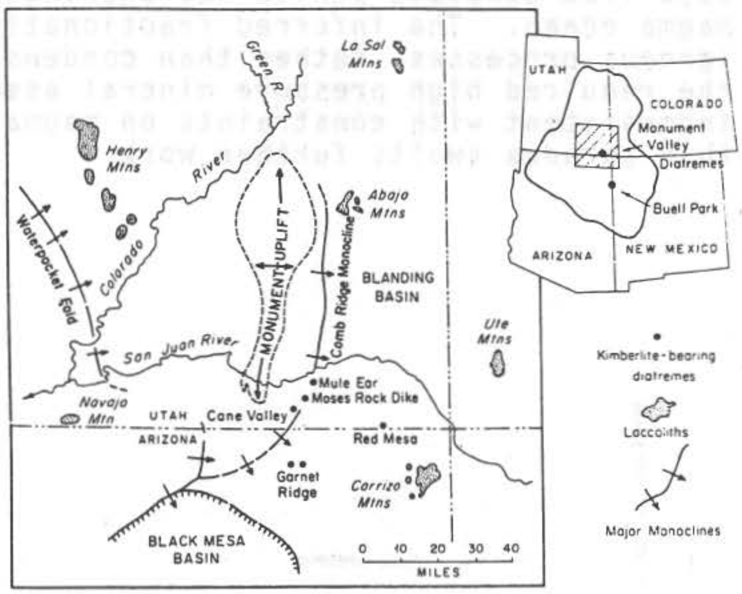


Figure 2

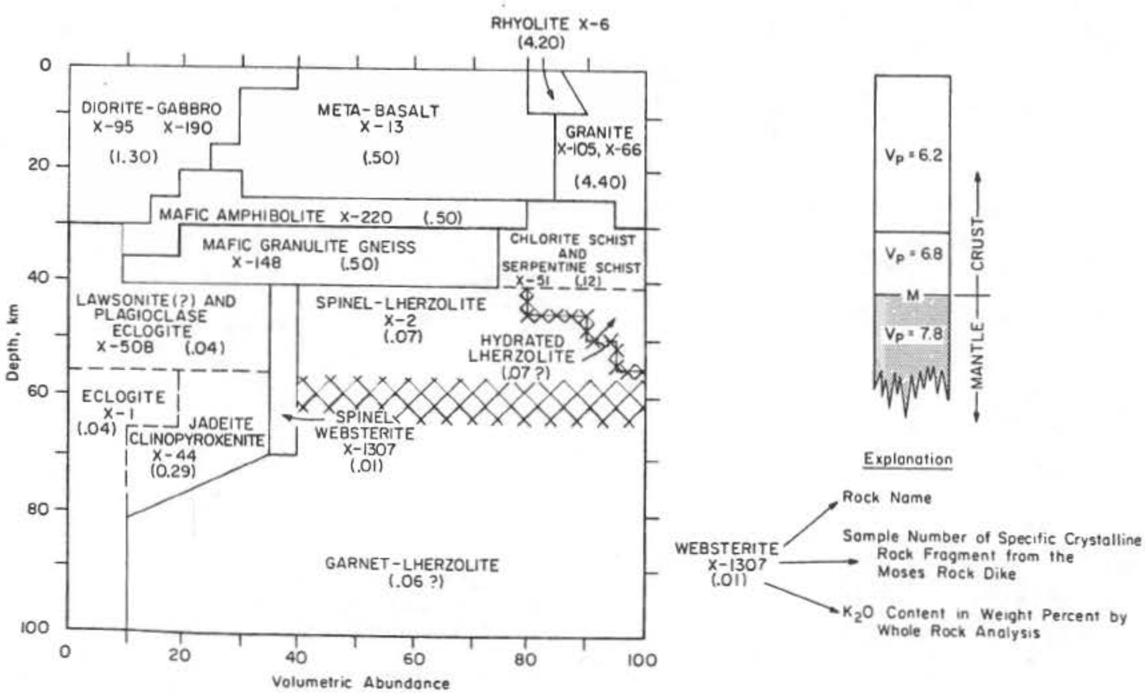


Figure 3

IGNEOUS EVOLUTION OF THE LUNAR HIGHLANDS, Gordon McKay, Mail Code SN6, NASA/Johnson Space Center, Houston, TX 77058

Current models successfully explain the general characteristics of highlands rocks in terms of magma ocean crystallization. Models of increased complexity are required to explain the details of highlands rocks. Requirements for the petrogenesis of highlands rocks constrain magma ocean compositions to be non-chondritic. Data from cumulate dunite and anorthosites support a fractionated magma ocean. The inferred fractionations are those expected for igneous processes, rather than condensation processes. However, the required high pressure mineral assemblages appear to be inconsistent with constraints on magma ocean depth. Solution of this paradox awaits further work.

THE BANDED ZONE OF THE STILLWATER COMPLEX: STRATIGRAPHY, PETROLOGY AND A MODEL FOR THE GENERATION OF ANORTHOSITE ZONES. L.D. Raedeke and I.S. McCallum, Department of Geological Sciences, University of Washington, Seattle, Washington 98195.

The plagioclase-rich rocks of the Stillwater Complex, i.e. the Banded zone, have been subdivided into three zones and twelve subzones. A detailed stratigraphic sequence has been determined for the entire section in the Contact Mountain area (4468 meters) and for a partial section in the Picket Pin area (935 meters). The relative proportions of cumulus minerals and whole rock modes have been determined as a function of stratigraphic height. Whereas the major zones show a fair degree of lateral continuity in terms of thickness and lithologies, many of the subzones and members within these subzones show significant lateral variations in thickness, modes, and textures. A simplified stratigraphic section, a brief description of the major units, and plagioclase and olivine compositions are presented in figure 1.

The dominant lithologic units in the Lower Banded zone (LBZ) are norite and gabbronorite with minor anorthosite, troctolite and gabbro members. Repetitive cycles of magmatic sedimentation have been recognized within the lower olivine-bearing subzone (OBZ I). Throughout the Middle Banded zone (MBZ) anorthosites are dominant; plagioclase comprises 82% (by volume) of this zone. Two complex olivine-bearing subzones are sandwiched between the thick anorthosites; attempts to identify repetitive cycles within these subzones have only been partially successful. The Upper Banded zone (UBZ) is composed of a lower olivine-bearing subzone and an upper subzone of uniform gabbronorite showing planar lamination. Although thick sections of the Banded zone are characterized by isomodal layering, the occurrence of modally graded layers, inch-scale layering, disturbed layering, cross-bedding, cut-and-fill structures, and discordant contacts attest to the action of currents during crystallization.

The primary processes operating during the formation of the Banded zone were fractional crystallization, multiple injections of fresh magma, magma mixing and convection and/or density currents. These processes are interdependent and each cannot be considered in isolation. Fractional crystallization was an important, even dominant, process during the formation of the Lower and Upper Banded zones. In general, the sequence and proportions of minerals precipitated and the systematic variation of plagioclase and pyroxene compositions in these zones (see figure 1) are compatible with those predicted from phase equilibria. Superimposed on this pattern are subzones in which olivine reappears as a cumulus mineral, numerous stratigraphic intervals composed of relatively thin anorthosites, and intervals in which the minerals deviate significantly from cotectic proportions. While the bulk of the LBZ and MBZ shows isomodal layering, modally graded layers are not uncommon. The data are strongly suggestive of the accumulation of crystals at the bottom of the magma chamber during the formation of the LBZ and UBZ. This does not necessarily imply that plagioclase settled through the magma; a more likely process is the deposition of plagioclase and mafic minerals by density currents that periodically swept across the floor of the magma chamber.

The stratigraphic sequence in the Middle Banded zone (MBZ) presents a major problem in interpretation. Plagioclase comprises ∿82% (by volume) of this zone. Since there is no known magma with a composition capable of producing rocks with this amount of plagioclase, we are forced to conclude that plagioclase has been significantly enriched within the MBZ. The following

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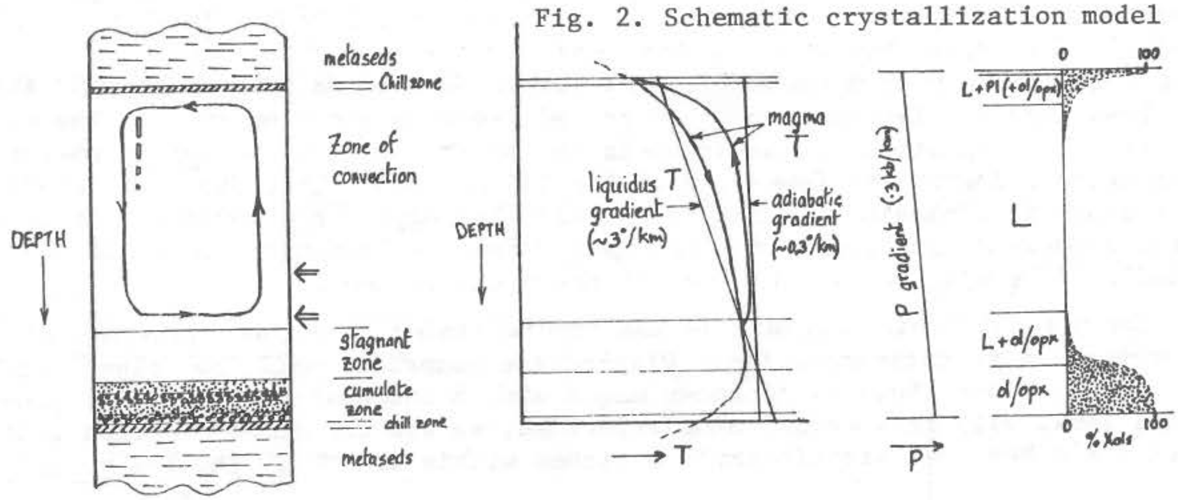
questions arise: (1) What is the source of the excess plagioclase? (2) Why did plagioclase accumulate at an intermediate level in the magma chamber? (3) What mechanisms were responsible for the accumulation of plagioclase. It is worth noting that the average grain size of plagioclase in anorthosites is ∿2X that in the two- and three-phase cumulates, that there is no systematic stratigraphic variations in plagioclase compositions within the MBZ, and that plagioclase has a density virtually identical to that of the coexisting magma. On the basis of these observations it appears that plagioclase in the MBZ remained in contact with the magma for a longer time than did plagioclase in other units and the system was fairly well mixed.

Mass balance calculations clearly point to the excess plagioclase as that which failed to accumulate on the floor of the magma chamber during the crystallization of the Ultramafic zone and the LBZ. It is worth noting that experiments on chilled marginal gabbro show both plagioclase and olivine to be on the liquidus at 1170°C and $fO_2=10^{-10}$. In an attempt to explain the data we have developed a model involving crystallization in a pressure gradient (figure 2). It is well known that pressure will depress the temperature of crystallization of plagioclase <u>relative</u> to that of mafic phases. It is conceivable that a magma with a composition close to the olivine-plagioclase cotectic may be saturated with plagioclase near the top of a thick magma chamber and saturated with olivine and/or pyroxene near the base. There would be little tendency for this plagioclase to settle (or float) insofar as its density was essentially the same as the melt. Furthermore, it is probable that the yield strength of the magma was sufficiently large to prevent differential crystal/liquid movement. Plagioclase would instead tend to "follow" the melt if, as seems likely, the system was convecting. Owing to heat loss through the floor and the superadiabatic liquidus T gradient, crystallization (presumably of olivine and/or pyroxene) was occurring at the base of the intrusion. It can be argued that the basal portion of the magma was not involved in the general pattern of convective circulation because of the increase in density produced by the crystallization of dense mafic minerals. The absence of size/density sorting and current structures in the ultramafic cumulates is consistent with this idea. We visualize a situation in which plagioclase crystallization was taking place in the upper, convecting part of the chamber while the Ultramafic zone was forming at the base. Plagioclase carried in convection currents might then be "deposited" (segregated) at some intermediate level in the magma chamber below which the liquid was denser than the plagioclase-liquid suspension whereas that above was less dense.

Fig. 2. Schematic crystallization model



THE BANDED ZONE OF THE STILLWATER COMPLEX

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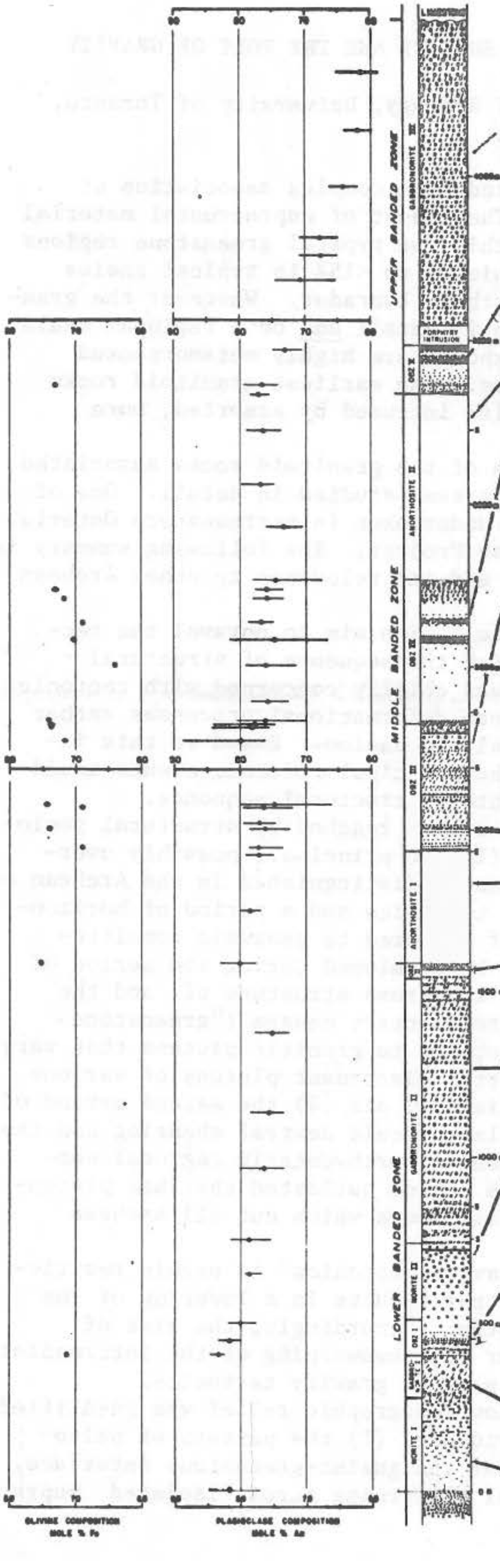

Figure 1: STRATIGRAPHIC SECTION

Gabbronorite III: Throughout this uniform subzone, plagioclase, augite and low-Ca pyroxene occur in approximately cotectic proportions. While planar lamination is almost always present, there is no preferred mineral orientation within the plane of layering. Below 3765 m orthopyroxene is clearly a cumulus mineral: Between 3765 and 3975 m, orthopyroxene occurs as poikilitic crystals containing abundant "inclusions" of small rounded augites. Above 3975 m, orthopyroxene occurs as poikilitic crystals containing numerous sets of oriented "001" augite exsolution lamellae. These sets of lamellae outline domains corresponding to original cumulus pigeonite crystals. Insofar as there is no significant change in mineral proportions associated with these textural changes, we conclude that low-Ca pyroxene formed as a cumulus mineral throughout this subzone, the poikilitic texture being the result of postcumulus recrystallization accompanying the inversion reaction.

Olivine-bearing subzone V (OBZ V): The basal member of this unit is a well-banded troctolite containing varied amounts of plagioclase, normally in excess of cotectic proportions. Modally graded layering, cross bedding and cut-and-fill structures are locally present indicating strong current action during the deposition of this troctolite.

Anorthosite II: This uniform unit is the thickest anorthosite in the Banded zone. Postcumulus augite and inverted pigeonite make up 10-12% of the rocks and disseminated sulfides are concentrated in two narrow layers near the base and top of this unit. The average grain size of plagioclase is approximately twice that of plagioclase in two- and three-phase cumulates.

Olivine-bearing subzone IV (OBZ IV): The lower boundary of this subzone is placed at the base of a well banded troctolite and the upper boundary at the base of the second thick anorthosite. The sequence troctolite-anorthosite-anorthositic troctolite-olivine gabbro (± gabbronorite) is repeated three times. The contact between troctolite and underlying olivine gabbro ranges from gradational to sharp, sinuous and discordant, whereas the upper contact between troctolites and overlying anorthosites is invariably sharp, planar and concordant. The uppermost troctolite is structureless, highly discordant, contains ameboidal olivine aggregates, and is virtually identical to the troctolite at the top of OBZ II.

Olivine-bearing subzone III (OBZ III): The predominant rock types in this 400 m of section are troctolite, anorthositic troctolite, anorthositic gabbro, olivine gabbro, and olivine gabbronorite. The various members are complexly interlayered, often on a centimeter scale. Olivine is present as a cumulus mineral through 80% of the subzone, but, except for eight narrow troctolite layers in which it comprises 40 modal percent, olivine abundances are low, making up between 1 and 10% of the cumulus mineral assemblage. The major units can be traced laterally but there are substantial lateral variations in the thickness, modes and textures of the thinner members. Two distinct cyclic sequences are repeated several times in the lower 275 m of this subzone: troctolite-anorthosite-anorthositic gabbro-olivine gabbro, and troctolite-anorthositic troctolite-olivine gabbro. Olivine gabbros and gabbronorites are usually isomodal with a well defined planar lamination. A distinctive feature of these rocks is the occurrence of polycrystalline aggregates of plagioclase (~2X2X1 cm), the long axes of which are aligned parallel to the lamination. By contrast, the anorthositic gabbros commonly show an erratic layering defined by alternating mafic and felsic layers. The upper 125 m of this subzone are characterized by a four-phase cumulate. In some samples, small embayed olivines are present in the cores of orthopyroxenes, clearly indicating a reaction relationship. In other samples, however, coexisting orthopyroxene and olivine do not show the reaction relationship and both appear to be cumulus minerals.

Anorthosite I: This subzone is composed entirely of a uniform plagioclase cumulate with postcumulus augite and inverted pigeonite. The average grain size of the plagioclase is coarser and more uniform than in the two- and three-phase cumulates above and below. Disseminated sulfides occur in the upper 8 m of this unit and reach a maximum concentration in the upper 2 meters.

Olivine-bearing subzone II (OBZ II): The upper 8 m of this subzone is a remarkable association of gabbro, troctolite, and gabbroic pegmatite. The contact between gabbro and overlying troctolite is sinuous and discordant, in contrast to the planar and conformable contact between the troctolite and overlying anorthosite. Irregular patches of gabbro are enclosed within troctolite and vice versa. Olivine in the troctolite occurs as large (>10 mm) "ameboidal" grians with inclusions of plagioclase set in a matrix of relatively small (1-2 mm) plagioclase crystals.

Gabbronorite II: Variable proportions of cumulus augite, orthopyroxene and plagioclase are seen in the lower section with four interlayered anorthosite members, two of which contain sulfides. In the central part of this subzone, mineral proportions are near cotectic and planar lamination is well developed. The upper 50 m are composed of five well developed but laterally discontinuous cyclic units showing modally graded layering defined by an upward increase in plagioclase/pyroxene ratio.

Norite II: Rhythmic inch-scale layering is well developed in the basal anorthositic norites. Upward in the section, modal proportions become more pyroxene-rich, approaching cotectic values, and layering becomes correspondingly more uniform (isomodal). Toward the upper part of the norite the habit of the cumulus orthopyroxene changes from subrounded to highly elongate. A distinctive, laterally extensive, mafic layer (2 m thick) and a complementary anorthosite layer occur ~20 m below the top of the subzone.

Olivine-bearing subzone I (OBZ I): The basal contact of this complex subzone is marked by the reappearance of cumulus olivine. The sequence troctolite-anorthosite-norite-gabbronorite is repeated five times with minor variations. The fourth troctolite contains abundant sulfides. The noritic members in the lower part of the subzone are characterized by a distinctive wispy banding; higher in the subzone layering in the norites becomes progressively more regular. This zone can be traced along strike for at least 20 km, but individual members are laterally variable, both in thickness and mode. Modal proportions are highly variable and generally non-cotectic.

Gabbronorite I: The lower contact of this unit is placed at the first appearance of cumulus augite. Proportions of augite, low-Ca pyroxene and plagioclase are largely non-cotectic with plagioclase increasing up section. Layering is poorly defined or isomodal. Pyroxenite inclusions are common in the upper 30 meters.

Norite I: Orthopyroxene and plagioclase are in approximate cotectic proportions in the lower uniform part of the unit, but plagioclase increases in abundance in the upper 100 m and layering becomes more pronounced. Layering is defined by alternating pyroxene-rich and plagioclase-rich layers with modally graded layering, scour-and-fill and slumping structures common near the top of the unit.

ARCHEAN SHIELD STRUCTURE AT THE SURFACE AND THE ROLE OF GRAVITY TECTONICS

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Archean shield terrain is characterized by a complex association of supracrustal rocks and granitoid rocks. The amount of supracrustal material is highly variable, ranging from > 50% within the typical greenstone regions of the Canadian, Australian and African shields to <15% in typical gneiss terrains such as western Greenland and northern Labrador. Wherever the granitoids-greenstone contact has been studied in detail and on a regional scale, it has been found that the oldest rocks exposed are highly metamorphosed greenstone or related supracrustal remnants. The earliest granitoid rocks are generally gneissic tonalite-granodiorite intruded by assorted, more massive granitoid plutons.

Until recently, the surface structure of the granitoid rocks associated with extensive masses of greenstone had not been studied in detail. One of the first regional structural studies was undertaken in northwestern Ontario, and formed part of the Superior Geotraverse Project. The following summary is largely based on results of that project, and its relevance to other Archean shields is uncertain.

Unlike conventional structural studies, which aim to unravel the tectonic history of a given region or establish the sequence of structural events, the Superior Geotraverse Project was chiefly concerned with tectonic mechanisms. Thus an insight was sought into deformational processes rather than reconstructing the detailed structural succession. Based on this insight and extensive geological mapping, the principal tectonic events could nevertheless be recognized and arranged into a structural sequence.

The following general conclusions have been reached by structural geologists of the Superior Geotraverse group: (1) Two principal, possibly overlapping periods of tectonic deformation can be distinguished in the Archean of northwestern Ontario, a period of gravity tectonics and a period of horizontal-motion tectonics; (2) large diapirs of foliated to gneissic tonalite-granodiorite with average diameters of 50 km developed during the period of gravity tectonics and are responsible for the gross structure of, and the major folds within, the metavolcanic-metasedimentary masses ("greenstone belts"); (3) late massive to foliated, dioritic to granitic plutons that vary from concordant, crescentic plutons to partly discordant plutons of various shapes and sizes were emplaced into the diapirs; and (4) the second period of tectonic deformation is characterized by large-scale dextral shearing and the development of major transcurrent faults under northwesterly regional compression. The strike-slip motions of this period outlasted the late plutonism, and led to the development of mylonitic zones which cut all Archean granitoid plutons.

Following Hans Ramberg, the term "gravity tectonics" is herein restricted to the formation of large structures that results in a lowering of the total gravity potential of a tectonic system. Accordingly, the rise of evenly spaced, major gneiss diapirs and/or the downwarping of the intermediate supracrustal rocks (greenstone belts) is genuine gravity tectonics.

Diapirism in this shield area with low topographic relief was identified by utilizing three lines of structural evidence, (I) the pattern of paleostrain within the gneiss, (II) the shape of the gneiss-greenstone interface, and (III) the magnitude of bulk horizontal shortening across isolated supracrustal masses (greenstone belts).

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An important element of (I) is horizontal flattening fabrics in the crestal region of dome structures. Such horizontal fabrics occur in the crestal region of diapirs and in high-viscosity layers folded under lateral compression (buckling). The possibility of buckle folding can be ruled out if the strain fabric has been imposed during "doming", and if the ratio of equivalent viscosity between the granitoid and supracrustal rocks is < 50. The large magnitudes of flattening characteristic for oval lower-order diapirs are compatible with model data if the lower-order diapirs are assumed to emerge from higher-order diapiric ridges. This assumption is born out by model experiments and the elongate shape of the natural higher-order diapirs in northwestern Ontario.

The shape of the gneiss-greenstone interface (II) in major diapirs depends partly on the nature and geometric pattern of horizontal density variations in the overburden. Diapiric ridges are obtained if the density of the overburden changes abruptly across a given horizontal line. Nappe-like diapirs develop if this horizontal line is relatively close to the base of the overburden, and if it involves a half-layer whose density is lower than that of the diapiric member. If the half-layer is exposed at the erosion level of natural higher-order diapirs, then it must correspond with a nappe-like geometry of the diapiric contact. The Rainy Lake diapir, first recognized by Lawson, is an example of such an asymmetric diapir.

The magnitude of bulk horizontal shortening (III) was crudely estimated for the Kakagi Lake region, east of Lake of the Woods. At an erosion level near the inflection lines of the diapiric contact, the magnitude of bulk horizontal shortening of interdiapiric model synclines is 25-35% of the mean horizontal breadth of the adjacent immature diapirs. The bulk shortening of the Kakagi Lake greenstone domain is about 30%, which suggests that all of the strain in the supracrustal rocks is the result of gravity tectonics.

Contacts between gneissic tonalite-granodiorite and greenstone show that the main tectonic deformation involved little if any ductility contrast. Evidence of large ductility contrasts during early tectonism is preserved, such as (i) dikes of gneissic tonalite-granodiorite within greenstone and (ii) angular greenstone inclusions in gneissic tonalite. Because the gneissosity was acquired during diapirism and while the ductility contrast was negligible, the early high ductility contrast was probably associated with the intrusion of tonalitic magma.

The original tonalite-granodiorite was either emplaced as discrete tabular batholiths, or formed a continuous layer below the extensive volcanic assemblage. As shown in Ramberg's experiments, tabular batholiths will spawn at least two gneiss diapirs. To estimate the minimum size of such possible batholiths, the size of the first-order diapirs must be determined. Because some of the larger elliptical diapirs are coalescent, they may be second-order structures within extensive diapiric ridges. Alternatively, the coalescence may be due to mushroom-like expansion of neighbouring mature diapirs. We studied the strain pattern in a coalescent zone between two gneiss diapirs whose diameters are about 100 km. The natural strain field is incompatible with mushrooming, and indicates clearly that the two elliptical diapirs are second-order structures. Thus the corresponding original batholith would have to be much wider than 200 km, the minimum length of the first-order diapiric ridge.

To date, not a single early batholith has been delineated in northwestern Ontario. Were it not for the Quetico and northern English River sedimentary belts as well as elongate greenstone masses like the Kenora-Savant Lake belt (Ontario Geological Map, West Central Sheet), the distribution of major ellipsoidal diapirs would be relatively uniform. In spite of

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an east-west alignment of the major axes of ellipsoidal diapirs, such uniform distribution is apparent on the adjacent Northwest Sheet of the Ontario Geological Map. It resembles the prominent salt dome pattern of the Gulf Coast region, and suggests that the buoyant tonalite-granodiorite material was confined to evenly and closely spaced tabular to elongate batholiths, or formed a continuous light layer below the heavy metavolcanic cover. Geophysical surveys support such a continuous layer, which could be due to widespread melting and/or tonalitization of greenstone. In South Africa, Carl Anhaeusser has found compelling evidence for widespread tonalitization and partial melting of supracrustal rocks adjacent to the Barberton greenstone belt.

The average density of metasedimentary rocks in the Quetico belt is close to that of gneissic tonalite. Within the belt, an array of magmatic diapirs with diameters of < 20 km has been recognized north of Thunder Bay. Apparently, the diapiric material was a crystal mush generated by widespread melting at depth. The present diapiric plutons contain large feldspar megacrysts whose strain fabric corresponds to the deformation pattern in model diapirs. Small magmatic diapirs are also present within the adjacent Shebandowan greenstone belt. They are probably common within other greenstone belts as well, and may be coeval with the large plutons of diorite-granite which cut the structural pattern of the gneiss diapirs.

The east-west elongation of most large structures which are shown on the Ontario Geological Map is due, in part, to horizontal-motion tectonics. The strain pattern of lower-order gneiss diapirs indicates that their elliptical shape is primary. So are the links of tonalitic gneiss between some higher-order diapirs that represent culminations on easterly trending ridges. Analogous structures have been obtained by diapiric modelling under conditions of crustal extension and normal faulting of the uppermost crust.

Evidence for large age differences between lithologically inseparable gneisses is accumulating, suggesting similar age differences between the (undated) greenstone enclaves in a given gneiss terrain. Perhaps we are seeing the net result of cyclic gravity tectonics resulting in composite gneiss domes and ridges which have experienced spasmodic diapirism and have intermittently born a heavy volcanic cover.

SYNTHESIS OF LUNAR HIGHLANDS PETROLOGY, GEOCHEMISTRY AND GEOCHRONOLOGY;
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Prologue - Among the terrestrial planets the moon represents the least complex example of an evolved body. The very simplicity of the moon's history makes it a facinating object for study. In fact, in many cases it is possible to pose a problem and perform an experiment on a limited number of samples and have some confidence in the generality of the observations from those few samples. The essence of petrologic geochemical and isotopic studies of the moon comes from the fact that the planet virtually lacks any H_2O or CO_2 or any evidence for tectonic processes other than those driven by impact. Chemical fractionation on the moon is assumed to take place almost exclusively by igneous processes (crystal settling and partial melting). The lack of water or an atmosphere precludes fractionation by weathering and sedimentation. Also the lack of water probably precludes the kinds of chemical migration often referred to occur in terrestrial metamorphic terrains. Although impact induced fractionation by volitization has been suggested it has not been unequivocally demonstrated to operate in any environment other than in the upper few meters of lunar soil, and is not relavent to the kilometer and larger scale impacts which dominate the moon's cratering record.

Regional variations - The lunar surface consists of (1) a light colored, heavily cratered, low density terrain termed the highlands and (2) a darker smoother, less cratered terrain confined to topographic depressions and termed mare. It is in the highlands that we have direct evidence (i.e. samples) of the early lunar crust, and its impact battered derivatives. However, the mare lithologies, mostly iron rich basalts, play a critical role in understanding the moon's early fractionation since they appear to be partial melts of which are chemically complimentary ultramafic cumulates, to the feldspathic highlands rocks, and isotopic evidence suggests that the source regions of the mare basalts evolved at the same time as the highlands. The tables summarize the compositions of key lunar rock types.

Bulk composition of the highlands - Lunar samples studies complimented by orbital measurements of composition suggest that the highlands composition is that of an anorthositic norite with about 25-28% Al_2O_3. Highlands rocks generally contain An_{80-97} plagioclase, low calcium pyroxenes with 45% or more enstatite component, magnesian olivine, subordinate magnesian augites, and in generally decreasing order of abundance minor amounts of Fe-Ti oxides, silica minerals, magnesian spinels, iron rich spinels including chromite, phosphates, troilite and metal. Figure 1 shows a range of rock compositions plotted on a plagioclase-olivine-silica liquidus diagram.

The composition of the highlands varies laterally and is less aluminous and richer in K,Th and by inference all LIL elements in the central nearside highlands sampled by the Apollo mission, than in the nearly two km higher region on the moon's far side. The lithophile-rich norite composition referred to as KREEP (Figure 2 and tables) is not as abundant on the far side, and studies of the Soviet Luna samples from the moon's eastern limb demonstrate the general lack of such materials at the surface. The book by S.R. Taylor (1) provides an excellent summary of both the arguments about the bulk composition of the lunar crust and its regional variations. Sample evidence for lateral heterogeniety in the moon's bulk composition are given by Jovanovic and Reed(2).

The vertical variations in the lunar crust is not well known; seismic evidence shows progressive increase in velocity with depth to about 25 km, suggesting decreasing effects of impacts. Lunar rock composition and the planets low gravity are not favorable for geobarometry. The geochemical scenarios for the moon typically assume a 200-300 km or more layered sequence,

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with feldspathic lithologies at the top, ultra mafic cumulates at the bottom, and residual liquid concentrated in the middle. Three categories of rocks were returned from the lunar highlands. Most (over 90%) sample consist of impact breccias many of which are impact melt rocks formed during the intense early cratering history of the lunar surface. Modeling of the cratering process by (3) suggest that the average depth of ejecta and the average depth penetrated by impacts is about 2 km with local penetrations to 15 km or more in the huge basin forming events. The characteristics of the breccias are: (1) they display clasts set in a fine grained matrix (2) the clasts are polymict (3) the bulk compositions look like mixtures not volcanic compositions (4) the clasts are more refractory than the matrix when the matrix is crystallized melt (5) most of the breccias yield ages between 3.84 and 4.05 billion years (6) almost all have more trace siderophile elements (typically Au over 1 ppb) than the mare basalts. A few samples form a second type of impact breccia, they are granulitic textured. Their characteristics are: (1) textures are either fine-grained mosaic or medium-grained poikiloblastic (2) chemically appear to be a mixture of plutonic rocks with almost no highland basalt (called KREEP and containing enriched trace lithophile elements characterized by light-enriched rare-earth elements) (3) impactites as suggested by clast-matrix structure, refractory clasts and enriched trace siderophile elements (4) mineral chemistry suggests derivation from a mixture of lunar plutonic rocks (5) isotopic age is skimpy and appears to be 4.3 - 4.1 AE. Some samples from the highlands are plutonic rocks--many of which are interpreted as being cumulates. The characteristics of lunar plutonic rocks are: (1) coarse-grained (relative to other lunar samples) - 1 mm to several cm (2) most display crushed and/or annealed (granulitic) textures. Those few others display igneous textures including oscillatory zoning in plagioclase and complex exsolution in pyroxene (3) some display isotopic ages near 4.5 AE. Others show ages as young as 3.9 AE and are believed reset by meteorite impact events. (4) Chemical compositions range from anorthositic through norite and troctolite to dunite (5) lithophile trace elements are low and unfractionated with respect to chondrites (6) trace siderophile elements are low (7) mineral compositions define several trends, at least one of which is magmatic.

Rocks which are texturally similar to lunar breccias are found in terrestrial impact structures, where it has been demonstrated that no chemical fractionation took place in 450 km^3 melt sheet (4). The amount of impact melt formed in a single event is a few times the mass of the meteorite (5) and at most a few percent of the volume excavated (6). Since over 1/3 of the highland rocks are melts, repeated episodes of impact are required to produce the observed populations.

The melt from the center of the impact becomes mixed by the turbulent motions set up by the impact and is much more homogeneous than the target. At the 65 km Manicouagan, Quebec structure material initially separated by km is mixed into individual 1 gm samples (4). The well mixed melt contains fine clastic debris from near the rim of the excavation. Since this fine grained debris is cold, it quenches the melt in seconds, initiating crystallization and giving the typical fine grained texture of impact melts (7).

The trace siderophile contamination of most lunar breccias by meteorite debris is accepted by most workers, and in fact low siderophiles have become a defining characteristic of endogenetically produced igneous rocks, e.g. the pristine rocks of Warren and Wasson (8). All of the rocks containing clasts set in a matrix studied by this author which have been analyzed, do in fact have high siderophiles. However, it has been recently argued by Delano and Ringwood (9) that the coarse grained plutonic rocks with low siderophiles are impact melts stripped of the siderophiles by the fractional crystallization of metallic iron from huge pools of melt. Delano and Ringwood further argue for a higher level of lunar refractory siderophiles than the Chicago group in part to shore up their argument that the moon is derived from the earth by a complex volatilation-fission mechanism. Anders (9) takes exception to their estimates for lunar siderophiles and argues for a distinctly different bulk composition and origin for the moon.

One inference from the trace siderophile contamination of the breccia is that the projectiles which bombarded the moon four billion years ago were different from those hitting the earth today (10).

The details of the timing of the lunar bombardment are significant for all of the terrestrial planets since as argued by Wetherill (11) all planets in the inner solar system are bombarded simultaneously, if not necessarily to the same intensity. Thus the cessation of bombardment about 3.9 AE implies the end of major bombardment of all the inner planets. If the granulitic impactites are considerably older than the other impact breccias it may be inferred that all the inner solar system was subjected to a post-accretion bombardment prior to its 4.05 - 3.85 AE age characteristic of most lunar impactites. Isotope systematics (by Rb-Sr, U-Pb, and Sm-Nd) suggest chemical and isotopic evolution of the lunar crust continued till about 4.4 - 4.2 AE. A small number of KREEP composition norites (Ages ∿3.95 AE) with low trace siderophiles and a lack of clasts have been identified that may also be derived by endogenetic igneous processes.

One focus of much of the current studies of the highlands is in refining the sequence of events shown schematically in Figure 3. Great interest exists in the nature and timing of the apparent ignous fractionation episodes prior to about 3.9 billion years ago. The Sr isotope and Nd isotope diagrams (Figure 4) clearly indicate that the mare basalts were not derived from a primitive mantle. Model dependent ages spanning the range from about 4.3 to 4.6 billion years have been proposed for various aspects of the process (12), and those ages are generally compatible with both crystallization ages for some of the plutonic rocks and model dependent ages for many highland lithologies. The formation of the dependent ages for rocks containing mixtures of a KREEP component argue for somewhat more recent fractionation than the mare and plutonic rocks (Lugmair and Carlson (13)).

The physical model commonly sited for the early fractionation is a globe encircling ocean of magma (1,14). In that ocean mafic cumulates formed at the bottom and feldspathic lithologies at the top. The mafic cumulates were subsequently partially melted to yield the mare basalts. However, no direct evidence exists that precludes the existance of much smaller sets of igneous intrusions.

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MARE BASALTS

	Low Ti Series						High Ti Series					
SAMPLE	Apollo 12 oliv-pig	Apollo 12 ilmenite	Apollo 12 feldspathic	Apollo 15 quartz	Apollo 15 olivine	Apollo 15 green gl	Apollo 11 high K	Apollo 11 low K	Apollo 17 Type A	Apollo 17 Type B	Apollo 17 Type C	Apollo 17 orange gl
SiO_2 %	46.42	44.44	46.56	47.94	44.90	45.2	40.37	40.37	39.48	38.70	38.33	38.8
TiO_2 %	2.14	4.59	3.31	1.87	2.41	1.1	11.77	10.18	11.77	12.28	12.26	8.8
Al_2O_3 %	9.15	9.81	12.53	9.49	8.81	15.1	8.84	10.40	9.38	9.26	8.66	6.4
Cr_2O_3 %	.52	.42	.27	.47	.57	.4	.36	.29	.39	.34	.59	.7
FeO %	21.33	20.77	17.99	20.76	22.41	13.7	19.28	18.73	18.84	19.23	18.10	22.2
MnO %	.29	.27	.27	.29	.30	.2	.23	.27	.28	.29	.27	.3
MgO %	9.48	8.55	6.71	8.55	10.41	12.14	7.56	6.92	7.97	7.62	10.17	17.4
CaO %	10.25	10.80	11.62	10.63	9.87	11.11	10.59	10.63	11.11	11.08	10.42	7.4
Na_2O %	.30	.31	.66	.32	.28	.4	.52	.40	.41	.38	.36	.4
K_2O %	.05	.07	.07	.06	.05	.1	.31	.07	.06	.05	.07	.08
P_2O_5 %	.08	.12	.14	.08	.08	.09	.17	.09	.05	.09	.05	.04
S %	.07	.09	.06	.07	.07	.06	.22	.16	.17	.16	.13	.07
Li ppm									9.9	8.6	8.8	10.7
Rb ppm	.8	.9	.48	.9	.7	.34	5.5	.74	.65	.39	1.20	1.11
Sr ppm	102.	143.	185.8	108.	95.		166.	174.	193.	136.	158.	209.
Ba ppm	54.	74.	120.	62.	48.	17.	308.	99.	83.3	64.5	68.6	76.4
Sc ppm							83.	90.	80.1	83.9	75.5	
La ppm	5.4	6.3	11.80	5.8	4.9	1.4	27.	10.	6.74	5.75	6.36	6.25
Ce ppm	14.	18.	29.1	15.	14.	3.8	76.	34.	24.0	18.6	22.0	19.0
Nd ppm	10.	15.	22.0	11.	10.	2.2	63.	31.	26.0	19.0	24.1	17.8
Sm ppm	3.4	5.9	7.57	3.6	3.2	.76	21.	12.	10.9	7.55	9.69	6.53
Eu ppm	.93	1.26	1.97	.97	.89	.21	2.2	1.9	2.16	1.53	1.81	1.80
Gd ppm	4.7	8.3	10.10	4.9	4.6	.91	27.	17.	16.7	11.8	15.0	8.52
Tb ppm			1.61			.15						
Dy ppm	5.2	9.9	9.73	5.5	4.8	1.1	33.	19.	19.2	13.5	17.0	9.40
Er ppm	3.0	5.5	5.0	3.3	2.7	.8	20.	12.	11.4	8.16	9.78	5.10
Yb ppm	2.3	5.3	4.8	2.7	2.2	.93	18.	11.	10.4	7.55	8.84	4.43
Lu ppm	.32	.74	.69	.33	.32	.14	2.6	1.6	1.43	1.09	1.25	.611
Zr ppm	89.	121.		95.	83.	22.						
Hf ppm						.42	18.	12.	9.6	6.9	8.7	
Th ppm						.08						
U ppm						.02					.16	
Ir ppb	.10	.07	.04	.006	.02	.22	.02	.006		.003		.214
Re ppb				.004	.002	.020			.019	.0015		.0553
Au ppb	.03	.08		.03	.026	.188	.03	.04	.029	.026	.19	1.07
Co ppm		60.	25.	42.	51.	72.	28.	16.	17.9	18.5	23.0	
Ni ppm	22.	7.		10.	45.	170.			1.5	1.		70.
Sb ppb				.02	.13	.12				.18		25.3
Ge ppb				5.7	9.4	37.			3.5	1.66		191.
Se ppb	170.	170.	120.	140.	150.	69.				176.		460.
Te ppb	30.	30.		2.7	3.0	3.3	8.	13.		2.1		49.
Ag ppb	.6	1.0		.90	1.	8.9	1.	2.		1.1		75.
Bi ppb	.3	.4		.16	1.3	.38				.099		1.53
Zn ppm	.8	.7		.14	1.5	19.	1.7	1.3	1.7	2.1	1.7	200.
Cd ppb		1.2		1.6	1.7	46.	6.	5.	1.9	1.8		260.
Tl ppb	.3	.35		.40	.32	1.13	.9	.3		.16		9.9
NORM (wt %)												
quartz			.80	1.01			1.93	2.67	1.07	1.17		
orthoclase	.30	.41	.41	.35	.30	.59	1.83	.41	.35	.30	.41	.47
albite	2.54	2.62	5.58	2.71	2.37	3.38	4.40	3.38	3.47	3.22	3.05	3.38
anorthite	23.47	25.17	31.02	24.28	22.64	39.11	20.87	26.38	23.58	23.41	21.81	15.43
diopside	22.58	23.27	21.72	23.53	21.66	12.70	25.24	21.40	25.64	25.49	23.98	17.13
hypersthene	40.71	35.64	33.60	43.69	33.80	23.18	22.47	24.46	22.60	21.72	23.37	13.79
olivine	5.40	3.50			13.74	17.70					2.41	34.48
ilmenite	4.06	8.72	6.29	3.55	4.58	2.09	22.35	19.33	22.35	23.32	23.28	16.71
apatite	.17	.26	.31	.17	.17	.20	.37	.20	.11	.20	.11	.09
chromite	.78	.62	.40	.69	.84	.59	.53	.43	.57	.50	.87	1.03

SYNTHESIS OF LUNAR HIGHLANDS PETROLOGY, GEOCHEMISTRY AND GEOCHRONOLOGY

Simonds, C.H.

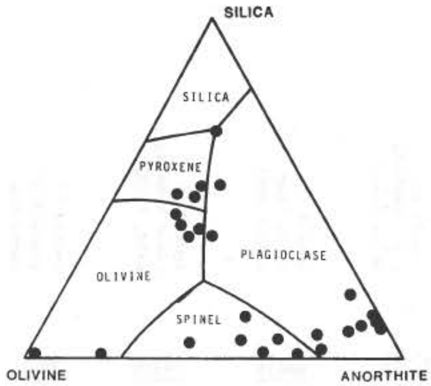

Fig. 1 Liquidous relations for pristine non-mare samples. Most clast laden impact melt rocks straddle the plagioclase-olivine cotectic, and extend throughout the less silicic parts of the plagioclase liquidous field.

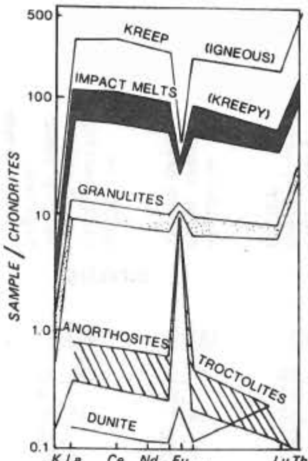

Fig. 2 Rare earths in a range of highlands compositions, most estimates for the bulk highlands correspond approximately to the field marked granulites.

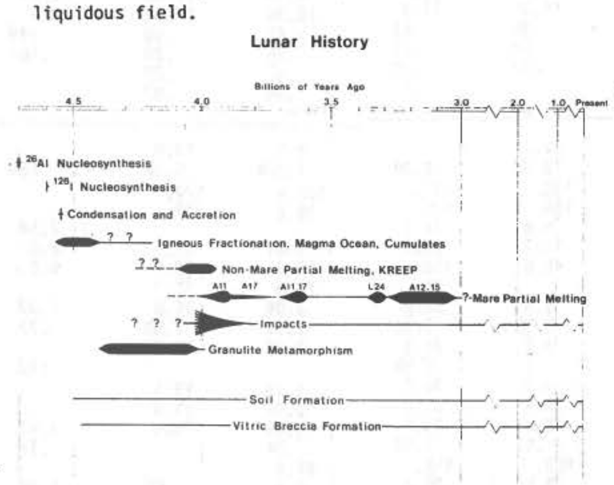

Fig. 3 Overview of lunar history.

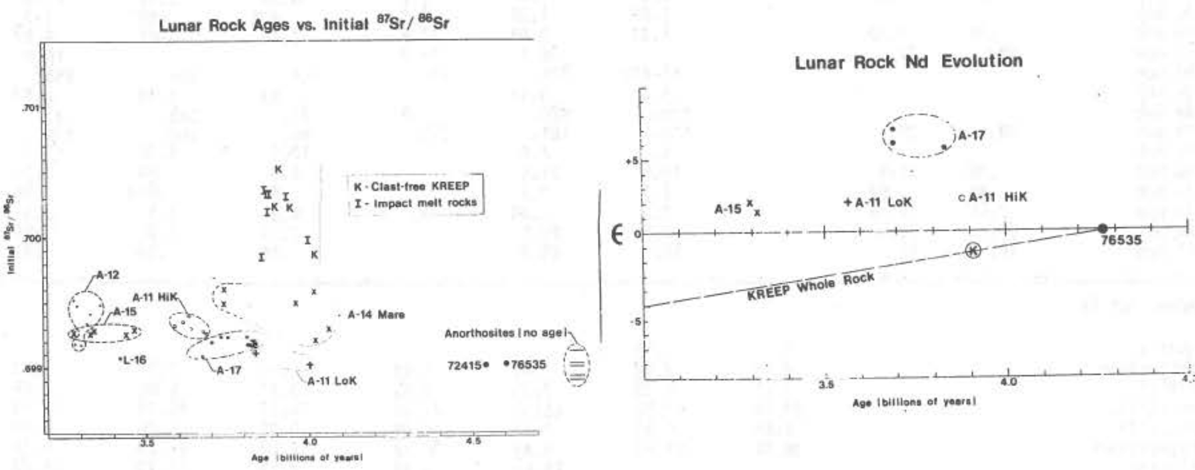

Fig. 4 A.-Sr evolution and B.-Nd evolution. K=Clast free KREEP basalts (15382, 15386), I=Impact melt rocks, 72415=a dunite, 76535=Feldspathic troctolite, A-11 LoK=Low K, High Ti mare basalts from Apollo 11, A-11 HiK are the high K, hight Ti Apolllo 11 mare basalts, A-12 and A-15 refer to Apollo 12 and 15 low Ti mare basalts.

SYNTHESIS OF LUNAR HIGHLANDS PETROLOGY, GEOCHEMISTRY AND GEOCHRONOLOGY

Simonds, C.H.

BRECCIAS

	Apollo 12 rock closest lunar sample to a granite, may involve liquid immiscibility		Apollo 14, clast laden breccia, very KREEPY	Apollo 14 clast free impact melt	Apollo 16 clast bearing impact melt, roughly lunar highland average composition	Apollo 16 clast bearing melt, KREEPY	Apollo 16 clast free impact melt	Apollo 17 clast bearing melt, KREEPY	Apollo 17 granulitic impactite
SAMPLE	12013 light	12C13 dark	14301	14310	61016	62235	68415	76315	79215
SiO_2 %			48.26	47.2	43.82	47.05	45.4	46.21	44.8
TiO_2 %	.3	3.3	2.06	1.24	.69	1.19	.32	1.50	.3
Al_2O_3 %	10.1	14.6	16.52	20.1	25.06	18.88	28.63	18.14	27.6
Cr_2O_3 %	.25	.162	.21	.18	.11	.19		.20	.098
FeO %	14.0	13.70	10.29	8.38	4.97	9.45	4.25	8.95	4.15
MnO %	.15	.166	.14	.11	.05	.13	.06	.12	.054
MgO %	8.50	9.0	9.98	7.87	10.48	10.0	4.38	12.02	5.84
CaO %	4.60	9.7	10.29	12.3	14.31	11.8	16.39	11.32	16.3
Na_2O %	1.12	1.45	.84	.63	.36	.42	.41	.60	.549
K_2O %	2.60	.40	.75	.49	.07	.33	.06	.26	.107
P_2O_5 %			.64	.34	.12	.39	.07	.29	
S %				.02		.1	.04	.97	
Li ppm					7.3		5.1	13.9	
Rb ppm	66.5	13.5	21.7	12.8	2.84	8.39	1.704	5.78	.489
Sr ppm			185.	188.	130.	152.	182.4	174.	
Ba ppm	3760.	390.	959.	617.	160.	530.	76.2	337.	
Sc ppm	25.	28.			6.6	16.5			7.14
La ppm	56.5	135.7	71.8	56.4	16.7	60.3	6.81	31.6	2.5
Ce ppm	151.	347.	201.	144.	46.0	160.	18.3	82.3	6.5
Nd ppm	74.	215.	121.	87.		94.	10.9	52.7	
Sm ppm	19.	59.	34.7	24.	6.9	25.2	3.09	14.8	1.03
Eu ppm	2.14	3.89	2.69	2.15	1.38	1.93	1.11	1.95	.77
Gd ppm			40.3	28.1	9.5	31.7	3.78	18.8	
Tb ppm	5.2	12.5			1.5	5.56			.23
Dy ppm	33.	75.	46.	32.7	9.7	35.0	4.18	19.1	
Er ppm			28.	19.1	4.8	21.0	2.57	11.4	
Yb ppm	30.	39.	25.	18.4	4.4	19.4	2.29	10.4	1.07
Lu ppm	4.2	5.2			.61	2.59	.34		.16
Zr ppm	690.	2070.		700.	209.	858.	97.5		
Hf ppm	27.	63.			4.9	20.7	2.4	.60	1.34
Th ppm	47.5	24.5		11.	1.6	7.83	1.26	5.69	.21
U ppm	14.2	10.4			.46	2.08	.32	2.52	.190
Ir ppb	.047	4.6		10.5	11.5	17.0	4.58	5.42	21.3
Re ppb				1.02	1.28	1.7	.434	.507	1.9
Au ppb	.21	3.10		4.31	9.55	17.6	2.65	3.21	8.27
Co ppm	24.	31.			36.7	54.2			18.8
Ni ppm				64-410	515.	830.	165.	256.	255.
Sb ppb				4.5	3.13		.53	1.49	2.79
Ge ppb				130.	620.	.8	73.	346.	33.
Se ppb	54.	99.		120.	181.	260.	98.	100.	176.
Te ppb				4.	4.8		13.5	4.04	17.0
Ag ppb	.88	31.4		10.8	21.4		4.8	.84	1.10
Bi ppb	.46	.63		2.5	7.3		.45	.098	.16
Zn ppm	2.06	4.10		2.3	.84	11.	4.8	3.1	2.3
Cd ppb	91.	34.		2.6	29.5		2.75	5.0	.98
Tl ppb	18.	12.		10.	45.3		.49	.31	.41
NORM (wt %)									
quartz			0	.16					
orthoclase			4.43	2.90	.41	1.95	.35	1.54	.63
albite			7.11	5.33	3.05	3.55	3.47	5.08	4.65
anorthite			39.10	50.57	66.56	48.66	76.11	46.04	72.53
diopside			6.68	6.91	2.98	6.18	3.92	6.80	6.74
hypersthene			36.74	29.61	6.63	32.72	12.62	25.20	4.31
olivine					18.68	3.38	2.75	11.19	10.22
ilmenite			3.91	2.36	1.31	2.26	.61	2.85	.57
apatite			1.40	.74	.26	.85	.15	.63	
chromite			.51	.27	.16	.28		.29	.14

SYNTHESIS OF LUNAR HIGHLANDS PETROLOGY, GEOCHEMISTRY AND GEOCHRONOLOGY

Simonds, C.H.

PRISTINE HIGHLANDS

SAMPLE	Anorthosite (60025)	Dunite (72415)	Feldpathic Troc (76535)	KREEP (15386)	KREEP (15382
SiO_2 %	44.3	39.9	42.9	50.83	
TiO_2 %	.02	.03	.05	2.23	2.17
Al_2O_3 %	35.2	1.5	20.7	14.77	14.9
Cr_2O_3 %	.03	.3	.1	.35	
FeO %	.5	11.3	5.0	10.55	9.2
MnO %	.02	.1	.07	.16	
MgO %	.2	43.6	19.1	8.17	7.4
CaO %	19.2	1.1	11.4	9.71	7.1
Na_2O %	.5	<.02	.2	.73	.85
K_2O %	.03	.0	.03	.67	.73
P_2O_5 %	.003	.04	.03	.70	
S %		0.01			
Li ppm		4.9	3.0	27.2	
Rb ppm	<.1	.066	.24	18.5	16.1
Sr ppm		2.24	114.	187.	195.
Ba ppm		3.27	32.7	837.	793.
Sc ppm					
La ppm	.28	.05	1.51	83.5	79.5
Ce ppm	.65	.07	3.81	211.	212.
Nd ppm	.42	.07	2.30	131.	127.
Sm ppm	.072	.022	.61	37.52	35.2
Eu ppm	1.04	.016	.73	2.72	2.77
Gd ppm		.030	.73	45.4	42.9
Tb ppm					
Dy ppm	.19	.035	.60	46.3	45.7
Er ppm	.05	.04	.53	27.3	28.1
Yb ppm	.048	.045	.56	24.4	24.0
Lu ppm	.006	.008	.079		3.43
Zr ppm		3.0	24.	970.	
Hf ppm	.02	.015	.52		
Th ppm			.16	10.	
U ppm		<.005	.056	.061	3.72
Ir ppb	.0057	.0052	.0054	.22	.0132
Re ppb	.0016	.0048	.0012		.0089
Au ppb	.0074	.255	.0025		.0033
Co ppm				23.	
Ni ppm	.3	149.	44.	12.5	18.
Sb ppb	.035	.47	.014		.17
Ge ppb	2.30	29.8	1.70	61.	47.2
Se ppb	21.7	4.9	4.1		72.
Te ppb	65.	<.36	.28		1.0
Ag ppb	.22	.25	.12		.44
Bi ppb	3.58	.41	.037		.29
Zn ppm	.17	2.1	1.2	3.5	2.6
Cd ppb	7.25	.37	.60	10.	86.6
Tl ppb	26.	.049	.12		3.2
NORM (wt %)					
quartz	NE= .07			7.62	
orthoclase	.18	0	.18	3.96	
albite	4.11	0	1.69	6.18	
anorthite	93.72	4.09	55.50	35.50	
diopside	1.30	.90	.70	7.39	
hypersthene		2.02	3.64	32.40	
olivine	.55	90.28	37.56		
ilmenite	.04	.06	.09	4.24	
apatite		.09	.07	1.53	
chromite	.04	.44	.15	.52	

ARCHEAN ROCKS IN THE SOUTHERN PART OF THE CANADIAN SHIELD

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Archean rocks in the southern part of the Canadian Shield compose two crustal segments of widely different age that largely evolved separately. Greenstone-granite complexes, 2,750-2,650 m.y. old, which are typical of much of the Superior structural province, are bounded on the south by older migmatitic gneisses and amphibolite that are in part 3,500 m.y. or more old. The greenstone-granite complexes have been called the greenstone terrane; the gneiss and amphibolite assemblages have been termed the gneiss terrane (Morey and Sims, 1976).

The greenstone-granite complexes consist of low grade mafic to felsic volcanics and clastic and chemical sedimentary rocks that are intruded by abundant granite-tonalite. They were formed during a time span of 50 to 100 m.y. adjacent to the pre-existing gneiss terrane, which 2,700 m.y. ago was part of a sialic protocontinent of moderate size. Although accumulation of the bedded rocks overlapped onto the older gneisses at least a short distance, there is no direct evidence that the majority of these rocks were deposited on a sialic crust. The environment of deposition possibly was analagous on a smaller scale to that of modern volcanic arcs and back-arc basins. The late Archean greenstone terrane probably was welded to the sialic protocontinent by late Archean plutonic magmatism and tectonic thickening of rocks in the boundary zone between the two crustal segments. In this zone, ∿2,700 m.y. granite-tonalite was deformed together with the bedded rocks in late Archean time, and the gneissic structure trends parallel to the boundary. The greenstone terrane stabilized tectonically at the end of the Archean.

Compared with the greenstone terrane, knowledge of the gneiss terrane is meager because of limited exposures, high grade of metamorphism, and multiple deformations, which have severely disturbed isotopic systems. Intensive studies in the Minnesota River Valley, summarized by Goldich and Wooden (1978), indicate that the migmatitic gneisses were formed over an interval of about 500 m.y. (3,500 m.y. - 3,000 m.y.) and consist of rocks of both igneous and sedimentary derivation that were folded and metamorphosed at least twice in the Archean and intruded by a major granitic pluton about 2,600 m.y. old. Subsequently, (∿1,800 m.y.) the isotopic systems were disturbed by a thermal event. The older rocks are tonalitic and granodioritic gneisses; the younger rocks are mainly adamellite. Scattered data from northern Michigan and Wisconsin are consistent with that from the MRV but less comprehensive. A tonalite gneiss from Watersmeet, Michigan, has a minimum age of 3,400 m.y. and probably is much older (Peterman and others, in press). Data on gneisses from other localities in northern Michigan and Wisconsin, mainly by W. R. Van Schmus and colleagues, indicate ages of ∿2,800 m.y.; but data from rocks in some localities indicate a previous crustal history. The long history of events during the Archean and subsequent tectonism in early Proterozoic time (Sims, 1976) indicates that the gneiss terrane was mobile, at least intermittently, from about 3,500 m.y. to 1,600 m.y., when it stabilized tectonically.

Representative analyses of rocks of Early and late Archean age from Michigan and Wisconsin are given in table that follows. The informally named granite near Thayer and the gneiss near Morse, Wisconsin, are interpreted as deformed facies of the Puritan Quartz Monzonite, which composes a large batholith in the greenstone terrane (Sims and others, 1977).

ARCHEAN ROCKS IN SOUTHERN PART OF THE CANADIAN SHIELD

Sims, P. K. and Peterman, Z. E.

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ARCHEAN ROCKS IN SOUTHERN PART OF THE CANADIAN SHIELD

Sims, P. K. and Peterman, Z. E.

Table--Representative Analyses

	(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)
SiO_2	68.6	62.2	67.8	69.3	72.4	74.1	71.0	62.0
Al_2O_3	15.2	16.3	15.1	15.2	14.6	14.2	13.5	15.8
Fe_2O_3	1.1	0.30	0.35	0.78	0.40	0.30	1.1	0.83
Fe	3.2	6.0	3.2	2.6	0.76	0.76	3.2	6.2
MgO	1.7	1.3	2.2	1.1	0.16	0.59	0.50	4.0
CaO	2.4	1.9	2.5	3.5	1.4	0.46	1.9	1.8
Na_2O	4.1	5.3	4.2	4.2	3.5	3.6	4.3	3.6
K_2O	2.7	4.1	1.8	1.7	4.8	5.2	4.0	2.6
H_2O-	0.02	0.02	0.04	0.13	0.02	0.05	0.07	0.18
H_2O+	0.87	0.65	1.2	0.55	0.50	0.64	0.84	1.8
TiO_2	0.74	0.58	0.41	0.35	0.11	0.21	0.09	0.59
P_2O_5	0.17	0.10	0.13	0.11	0.04	0.06	0.06	0.20
MnO	0.05	0.15	0.06	0.05	0.00	0.01	0.11	0.06
CO_2	0.08	0.52	0.05	0.08	0.08	0.07	0.12	0.00
SUM	101	99	99	100	99	100	101	100
Rb,ppm	184	96.3	51.4	59.2	102	136	119	58.5
Sr,ppm	194	64.8	304	277	94	46.6	180	337
U,ppm	1.8	6.8	1.1	0.7	20.5	17.6	7.4	2.8
Th,ppm	8.7	36.0	21.3	11.5	28.9	67.4	32.4	8.6

EARLY ARCHEAN:
(1) Tonalitic gneiss near Watersmeet (M-45L)
(2) Sheared facies of tonalitic gneiss near Watersmeet (M-48)

LATE ARCHEAN:
(3) Tonalitic gneiss of granite near Thayer (M-147-1A)
(4) Tonalitic gneiss near Morse, Wisc. (W-59)
(5) Granite of granite near Thayer (M-71)
(6) Leucogranite dike intruding gneiss at Watersmeet (M-45H)
(7) Puritan Quartz Monzonite (D1729)
(8) Metagraywacke (M-91)

IDENTIFICATION AND SIGNIFICANCE OF GEOCHEMICAL DISCONTINUITIES IN ARCHEAN VOLCANIC TERRAINS

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The concept of geochemical discontinuities within Archean volcanic successions provides a means of identifying and quantifying critical changes in petrogenetic processes during the development of early volcanic crust. Archean volcanic terrains are commonly made up of a succession of rock units from stratigraphically lower, predominantly basaltic volcanic rocks of tholeiitic affinity to overlying intermediate and felsic calc-alkaline rocks. Although this relationship cannot always be clearly demonstrated, the concept of geochemically more primitive volcanic rock types underlying geochemically evolved volcanic rock types appears to be valid.

Studies in greenstone terrains of the Superior Province of the Canadian Shield show that the transition from lower tholeiitic volcanics to overlying calc-alkaline volcanics, which occurs at about the mid-thickness in well developed successions, represents an important geochemical discontinuity characterised by the following features:

1) As a suite the calc-alkaline volcanics contain higher SiO_2 than the tholeiites
2) Al, alkalies and some LILE elements have higher abundances and greater variability in specimens of the calc-alkaline suite
3) The tholeiites show a greater range in Mg-number reflecting an iron enrichment trend not displayed by the calc-alkaline suite
4) Chondrite normalised REE abundance patterns in the calc-alkaline rocks show marked light/heavy fractionation (Ce_N/Yb_N 3.6 to 5.6) compared with the flat patterns observed in the tholeiite suite. Further, in the calc-alkaline suite there is a decrease in total REE abundance with increase in SiO_2 which is the reverse of that observed in the tholeiite suite.
5) Sc abundances are characteristically lower in the calc-alkaline rocks (20-30 ppm) compared with the tholeiitic rocks (30-45 ppm).

The differences in the REE abundances observed in the volcanic rocks on either side of this discontinuity provide a major constraint to any petrogenetic model. Model calculations assuming a chondritic mantle indicate that very small amounts of melting (<5%) of a spinel lherzolite assemblage leaving clinopyroxene in the residuum may explain the fractionated patterns of the calc-alkaline rocks whereas greater degrees of melting (10%) and subsequent shallow fractionation can explain the patterns observed in the tholeiites. Other geochemical features are consistent with this model. There is thus no compelling reason in this case to suggest that calc-alkaline rocks indicate a significantly different tectonic environment. However a second major geochemical discontinuity separating intermediate from felsic volcanic rocks toward the top of some relatively well preserved Superior Province greenstone belts and defined by a greater degree of light/heavy REE fractionation requires the existence of a geochemically evolved source.

A model consistent with observed compositional variations in Superior

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Province greenstone belts is that initially high degrees of melting of Archean mantle are followed by successively smaller amounts of melting and finally by reworking and melting of the lower portions of the volcanic pile as they impinge on the mantle P-T regime. From a tectonic point of view this is clearly different from a modern oceanic ridge situation in which continuous magma production produces spreading. Rather the model suggests a finite thermal event which produces a series of magma types as it decays.

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SIGNIFICANCE OF GRANULITE METAMORPHISM FOR STABILIZATION OF PLANETARY CRUST: CHARNOCKITE FORMATION AT KABBALDURGA, S. INDIA: ROLES OF CO_2 AND H_2O; SPECULATIONS ON VENUS AND MARS; J. V. Smith and R. C. Newton, Dept. of the Geophysical Sciences, University of Chicago, Chicago, Ill. 60637; A. S. Janardhan, Dept. of Geology, Mysore University, Mysore, India.

Stabilization of the Earth's crust is enhanced by loss of volatiles, thereby providing greater resistance to melting and deformation. We describe (i) conversion of amphibolitic gneiss to charnockite at Kabbaldurga, (ii) formation of granulite rocks by CO_2-rich solutions escaping from Archaean basaltic magmas, and (iii) speculations on crustal metamorphism in the inner planets.

Charnockite formation at Kabbaldurga (1). The development of the abundant suite of charnockite rocks (i.e. dark, green-to-brown, orthopyroxene-bearing granitoids) in Precambrian terrains is fundamental to evolution of the crust. Charnockite apparently derives from both igneous and sedimentary rocks (2) by metamorphism, in which associated pore fluids are low in H_2O because the associated opx-K-feldspar is limited to $P(H_2O)<700b$, being replaced by either biotite-quartz or melting at higher $P(H_2O)$. Primary CO_2-rich inclusions are characteristic of granulite (3) and charnockite (4), and CO_2 probably dilutes the H_2O.

The Kabbaldurga quarry will become classic for partial conversion of Archaean amphibolitic gneiss (~2700my) to charnockite. Diffuse patches and stringers of charnockite obliterate the gneissic foliation, and their textural positions strongly suggest local action of volatiles whose access is controlled by deformation. Rock analyses show little chemical change except for H_2O,CO_2 and perhaps K(5). The Kabbaldurga occurrence is one of many in S. Karnataka. These lie in the northern fringe of development of charnockite massifs in the south (6), and provide a key to the wholesale conversion of continental crust to granulite.

Optical petrography and electron microprobe analyses show that increasing alteration of acid gneisses causes green-brown discoloration of the feldspars, ultimate loss of hornblende and biotite with growth of orthopyroxene. Basic inclusions (probably relics of basaltic dykes) transform via symplectites to equigranular diopside-hornblende plagioclase rocks. The ubiquitous green veins which lace the feldspar consist of Fe-rich, Al-poor chlorite (SiO_2 33 Al_2O_3 11 total Fe as FeO 32.2 MnO 0.2 MgO 11.6 CaO 0.4 wt.%) and Mn-bearing calcite (e.g. 4.8% MnO). These veins account for the greenish color of the feldspars, and represent texturally the waning phase of metamorphism.

We conclude that (i) the granulite metamorphism involved localized action of volatiles low in H_2O (because of absence of melting), and by implication rich in CO_2, with no major change of P and T,(ii) charnockitization occurred at high level, probably at P = 3 to 5kb, (iii) the waning phases of metamorphism were characterized by cooler solutions, richer in H_2O, which were confined to localized channels and which deposited chlorite and Mn-calcite.

Thus we envisage copious streaming of CO_2-bearing solutions along pathways generated by brittle deformation, and go on to consider how granulites may be formed by action of CO_2-rich solutions ascending through the crust during Archaean time.

Formation of granulite rocks. Key aspects of the postulated charnockitization of the Archaean crust include (i) the source and vector of the carbonic solutions, and (ii) the possible P-T regime which would avoid melting or capture of CO_2 into scapolite.

Decarbonization of the mantle is the most likely source, but free CO_2

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cannot coexist with peridotite under any but extreme geothermal conditions, as in a rising plume, because of the high thermal stability of mantle carbonates (7). Basaltic magma would be generated in such an upwelling, and there should be insufficient carbonate to saturate the magma with CO_2 (8,p.119), so that basalt emplaced copiously into the lowermost crust is the likely CO_2 vector (9,p.248). During crystallization, nearly pure CO_2 is emitted while H_2O is retained quantitatively in the rest-magma at deep crustal pressures (8). The congealed basalt is a massive addition to the continental base, and, as geotherms steepen over the rising plume, partial remelting provides the voluminous tonalite magmas which rise to give higher-level Archaean terranes (10). Widespread granulite metamorphism and depletion of LIL-elements are contemporaneous (11).

Fig. 1 shows how CO_2 could be supplied by crystallizing basalt at the base of the continental crust with the Archaean geotherm lying between moderate (12) and extreme (13) limits still under controversy. The possible thickness of the crust is limited by the dry basalt solidus to 15-~40km. The passage of free CO_2 through a deep crust, either of basic or intermediate composition, is constrained by two curves for the stability of orthopyroxene & scapolite [of composition essentially 3 labradorite + $Ca(CO_3,SO_4)$] relative to plagioclase, diopside + quartz + CO_2. The two curves, labeled AN90 and AN50, respectively, were constructed from experimental (14) and thermodynamic (15) data. Streaming of CO_2 could have produced intermediate (tonalitic) scapolite-free granulites to within about 12km of the surface under a very steep geotherm, and to even higher levels in acid rocks with more sodic plagioclase. Charnockitization of the Kabbaldurga acid gneisses at ~600°C and 3-5kb would require only moderately steep geotherms. At depths less than several km on Earth, the process might not operate because of low temperatures (<500°C) and readily accessible H_2O to dilute the CO_2.

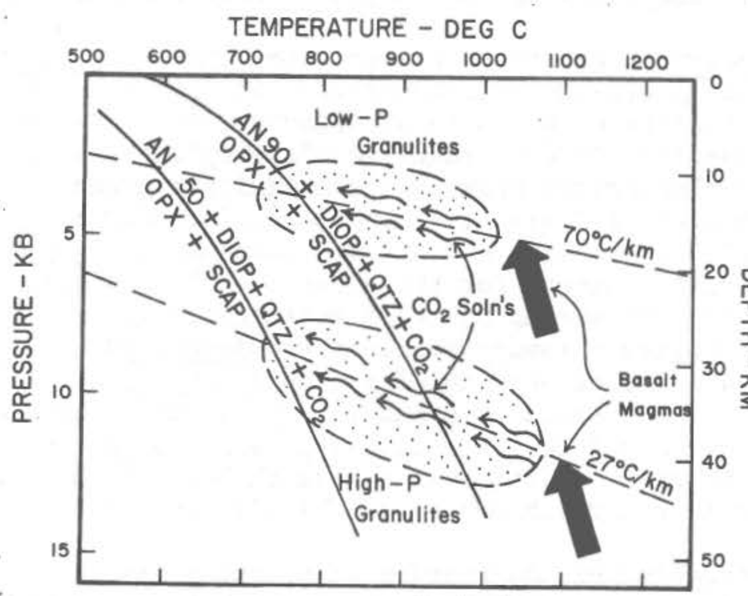

Fig. 1. Formation of granulite crust by ascending CO_2-rich solutions. Arrows suggest emplacement of continental crust during Archaean time. Depth of crust is limited by dry basalt solidus and either a moderate geothermal gradient (12) or a high one (13). Crystallization released CO_2 which streams upward and causes low-$P(H_2O)$ recrystallization. Calculated curves for the stability of the assemblage scapolite (meionite 65 with $CO_3/SO_4=1$) plus enstatite (opx) relative to the assemblage plagioclase (An90 or An50) + diopside + quartz + CO_2 provide limits to CO_2 migration. Under high geothermal gradients, CO_2 moves freely in the lower and middle crust. Under moderate gradients, CO_2 is absorbed in the deep crust to form scapolite in basic rocks but may effect charnockite formation in intermediate and acid rocks at mid-crustal levels.

Speculations on Mars and Venus. Fig. 2 summarizes pertinent phase equilibria and model pressure-temperature curves for Mars and Venus. For Mars, amphibole

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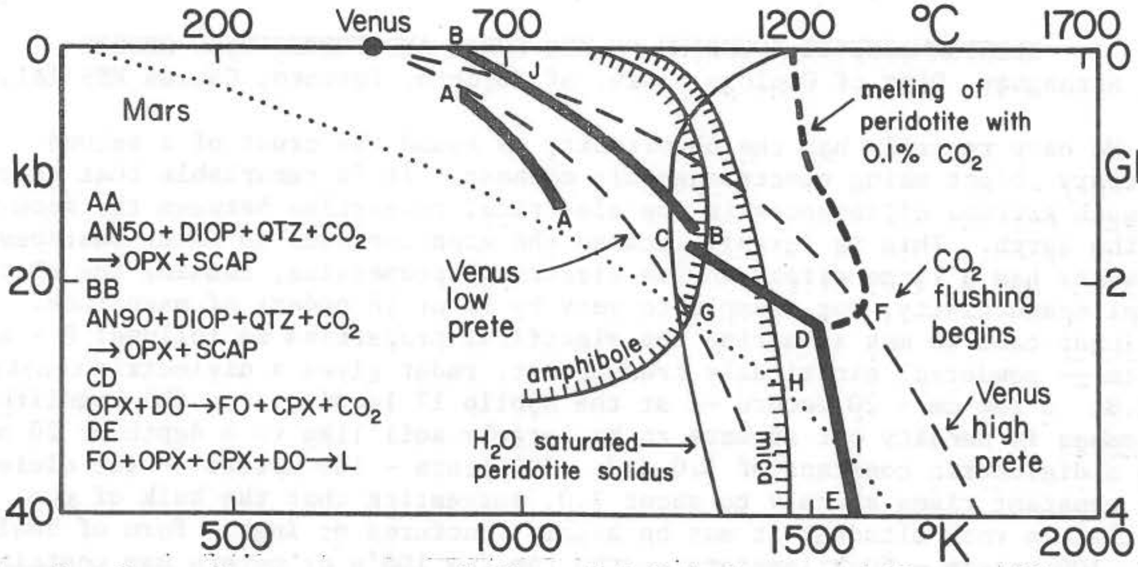

Fig. 2. Controls on storage of CO_2 and H_2O in Venus and Mars. All curves from (16), except AA, BB from Fig. 1. Dotted and dashed lines are estimated pressure-temperature curves for Mars and Venus. Generalized stability curves for phlogopite mica and amphibole are compared with the solidus for H_2O-saturated peridotite. Heavy lines show stability in presence of CO_2.

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and mica would become unstable at G and H (23 and 29kb) for the depicted P-T curve in peridotite with the same Fe/Mg ratio as Earth. If Mars is richer in Fe^{2+} than Earth, these depths are reduced. Storage of CO_2 in solid carbonate should cease at I(33kb), but trapped CO_2-rich magma could exist below I to give a low-velocity zone.

For Venus, we prefer to use the higher of the P,T curves, which is based on analogy with an oceanic geotherm rising from 470°C at the surface, and indeed earlier PT curves should be hotter. For the present curve, CO_2 would be liberated at all depths above F, and the released CO_2 would flush out H_2O from overlying rock by analogy with Kabbaldurga charnockite. Scapolite is restricted to a shallow layer above J for basic rocks, and cannot occur in an andesitic crust. This is consistent with the CO_2-rich atmosphere. A thickness of (say) 10km basalt with (say) 5% scapolite (~5 wt.% CO_2) corresponds to only ~25m of limestone, which is much less than the amount in the Earth, and much less than the CO_2-equivalent in the Venus atmosphere. Can volatiles be returned by subduction? No obvious mechanism is available for H_2O, and the implication is that Venus is essentially dry. Thus any primordial water must have lost its H through the atmosphere. For CO_2, subduction is possible for scapolite-bearing basalt, but may be less efficient than subduction of limestone on Earth. In general, the Venus crust should show the greatest development of granulite for any planet.

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ELECTROMAGNETIC SOUNDING OF THE LUNAR AND TERRESTRIAL CRUSTS

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We have recently had the opportunity to sound the crust of a second planetary object using electromagnetic methods. It is remarkable that there are such extreme differences in the electrical properties between the moon and the earth. This is largely because the moon contains no water whatsoever and water has a large effect on the electrical properties, causing the electrical conductivity, for example to vary by 17 or 18 orders of magnitude. In the lunar case we may summarize the electrical properties as follows: 0 - a few cm -- powdered, elctrically transparent, radar gives a dielectric constant of 2.8; a few cm - 20 meters -- at the Apollo 17 landing site the regolith increases in density but appears to be largely soil like to a depth of 20 m with a dielectric constant of 3.0-3.2; 20 meters - 100 meters -- the dielectric constant rises sharply to about 7.0, suggesting that the bulk of the material is rock although it may be highly fractured or in the form of boulders; 100 meters - 50 kilometers -- the top few 100's of meters may contain some electrical reflectors due to stratigraphic layers as revealed by the sounder experiment, but as indicated by all experiments the electrical resistivity is still very high. The surface layers are therefore water free and cool; 50 kilometers - 1000 kilometers -- from global sounding experiments the resistivity is seen to drop sharply, probably a consequence of increasing temperature. The apparent temperature inferred depends upon the composition of the material and we cannot make definite statements; 1000 kilometers - 1740 kilometers -- the electrical resistivity appears to reset again perhaps just due to further increasing temperature.

The electrical structure of the moon is quite straightforward, a simple drop in resistivity with depth as temperature increases and no effect of water.

In the earth, the situation is quite different. In sedimentary basins the resistivity of shales and other fluid filled sediments is very low. Large parts of the shield are covered with glacial lake clays and again the surface resistivity is locally very low. In tropical laterites the resistivity is also low, so that to study the crustal electrical properties one must choose the location carefully. In recent years a number of large scale resistivity soundings have been done by others and we have done a large number of magnetotelluric soundings in Precambrian crustal regions. Where the surface has been glaciated and where there is very little clay present,it is possible to sound the crust to depths of 20-30 kilometers even at frequencies of 1-10 hz. In a general way there is a thin near surface layer with a resistivity of a few 100 ohm-m. This rises rapidly to very high resistivity values, occasionally approaching as much as 100,000 ohm-m. This horizon must contain very little moisture and is highly transparent to electromagnetic energy. These rocks are fracture free. At depths of 10 - 20 kilometers, the resistivity drops very sharply and then rises again with depth. From very low frequency soundings (i.e. periods of hours) the resistivity is then observed to drop again. It appears that there is low electrical resistivity at depths of 50 - 100 kilometers presumably due to temperature increases as in the lunar case. We are left with trying to explain the presence of a low resistivity layer which is quite widespread at a depth of 10-20 kilometers. Could this be due to the presence of tiny amounts of trapped fluids and could it be related to the low velocity zone now reproted so commonly by seismologists even in the Canadian Shield?

THE ROLE OF WATER IN PLANETARY TECTONICS. Jeffrey L. Warner, SN NASA Johnson Space Center, Houston, Texas 77058

Introduction

Water might be the most important chemical constituent in a planet's mantle and crust in determining each planet's tectonic development. This is because a small difference in water content results in a large difference in physical properties. No other chemical constituent results in such basic changes in physical properties. For example, the difference between zero water (as is the case for the Moon) and 1% water (as might be the case for Earth) is equivalent to a difference of 200°C to 600°C in the temperature of the start of melting. This difference in melting temperature is reflected in many other properties such as differences in viscosity at the same temperatures and pressures.

These statements concerning water are made with full understanding of the effects of other chemical components on the bulk properties of planets. Parameters such as the metal:silicate ratio dominates the bulk density of a planet. But the metal forms a planetary core and does not have a strong effect on the subsequent development of the mantle and crust. The content of K, U, and Th have a controlling influence on the internal heat budget of a planet. However, a factor of 2 in the bulk abundance of radioisotopes has a surprising small effect on a planet's thermal history. This point is illustrated by the various thermal models calculated by Hubbard and Minear (1976) for the Moon.

The Role of Water

The major effect of one percent water in a planet's mantle is reduction of the temperature at which incipient melting occurs by 200°C to 600°C, with a consequent dramatic effect on the mantle's viscosity. The result is that a hydrous planet will, throughout its history, have more extensive regions where melt is present and where convection is happening. These statements are relative to an anhydrous planet. These effects are partly illustrated in Figure 1 where the results of simple (non-convecting) thermal model calculations are presented as an aid to understand the principle. The Figure shows the presence of zones where melt is present for a Moon-sized planet as a function of time for the cases of an anhydrous (dry) and hydrous (wet) planet. Johnston et al. (1974) considered the effect of water on the development of Mars and reached compatible conclusions.

These calculations are not intended to model the Moon or any other planet. They simply illustrate the dramatic effect that one percent water can have on the development and history of a planet.

Another way of looking at the role of water is how it effects the thickness of the lithosphere as a function of time. At any time in the development of two similar planets, except one is hydrous and one is anhydrous, the anhydrous planet will have a thicker lithosphere than the hydrous planet. Lithospheric thickness is a prime parameter in controlling the tectonics that is possible on a planet's surface. If the lithosphere is thin, it is possible to break into plates or some other type of structure. The broken parts of the lithosphere are then capable of transmitting lateral and vertical stresses. If the lithosphere is somewhat thicker, it will not break into plate-like parts.

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Only vertical stresses will be transmitted. Finally, if the lithosphere is very thick, neither lateral or vertical stresses will be transmitted. Kaula (1975), Solomon and Chaiken (1976), and Warner and Morrison (1978) have written about this dependence of tectonic styles on lithospheric thickness and referred to the first style as active tectonics, the second style as volcanic tectonics, and the final style as quiescence.

Does the presence of water in a planet effect the composition of the buoyant crust that forms on top of a planet-wide magma ocean? The Moon's crust is anorthositic because plagioclase is the only mineral that can crystallize from the magma ocean in abundance and has a density that is low enough so that it can float. But in a planet that is hydrous, the density of the magma ocean will be lower and it will be more difficult, perhaps impossible, for plagioclase to float. Is the tonalite in the Earth's crust a hydrous planet's equivalent to the anorthositic material in the crust of an anhydrous object like the Moon? What is the nature of the petrologic response to the lower density of a hydrous magma ocean? What is the nature of the fractionation that allows tonalite rather than anorthosite to be the early, buoyant crust of a planet?

There are several lines of observation and theory that suggest that the several planets each have a distinctive and different abundance of water.

Condensation From the Solar Nebula

Models for the thermodynamic condensation of dust from a nebula of solar composition, and the subsequent collection of those dust grains into planets, indicate that there should be a systematic difference in water content among the planets. The assemblage of minerals in the dust at any place in the nebula is a function of the temperature in that region of the nebula. The lower the temperature, the more volatile elements should be present in the dust: at high temperatures the dust will consist of anhydrous, refractory oxides; at moderate temperatures the dust will consist of anhydrous silicates; at low temperatures the dust will consist of hydrous silicates and other volatile-containing minerals; and at very low temperatures the dust will consist of ices of H_2O, CO_2, NH_3, and CH_4.

Temperature within the solar nebula decreases as distance from the sun increases. Since a planet is believed to accrete from dust in its vicinity, the position of the various planets should be an indication of each planet's content of volatile elements. Specifically, those planets near the sun should contain less water and other volatiles than the planets farther from the sun. There should be an increase in the abundance of water from Mercury to Venus to Earth to Mars. The Moon, which is completely anhydrous, represents a complication to this simple scheme.

Direct Observation

By direct observation, we know that Earth's atmosphere and lithosphere contain water, and that the lithosphere of the Moon is anhydrous. Note that the water molecules in the rusty spots on some Apollo 16 rocks have been demonstrated to be of terrestrial origin.

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The atmospheres (plus hydrospheres) of the terrestrial planets contain distinctive amounts of water. Expressed as atmospneres of water these values are:

Venus - .1 to .4 atmospheres water
Earth - 300 atmospheres water
Mars - .007 atmospheres water plus water in the polar caps and within the regolith

Hypothesis

Arguments, based on spacecraft observations of Venus, have been constructed that suggest that Venus contains less water than Earth. In fact, the arguments suggest that the mantle of Venus is anhydrous. Other arguments, based on spacecraft observations of Mars, have been constructed to show that Mars contains greater abundances of water than does Earth. These arguments are summarized below.

Water in the Earth

We really do not know how much water is contained in the Earth's mantle. Various probes into the Earth's mantle yield conflicting data. Granulite inclusions in basalt and kimberlite pipes are anhydrous, but kimberlites and basalts themselves contain water. If the Earth's mantle contains 1% water, than the entire mantle would contain the water equivalent of 20 oceans. Although we know the rate at which the Earth is degassing today, we do not know the rate that the Earth degassed in the past. We cannot answer the question "How degassed is the Earth?".

Water in Venus

Based on the similar density of Earth and Venus, the similar K/U ratio of Earth and Venus (Surkov, 1977), and preliminary data on the argon isotopic abundances as determined by Pioneer Venus and Venera 11 and 12, Warner (1979b) argued that for the temperature range at which most silicate material condenses from a nebula of solar composition, the condensation histories of Earth and Venus were similar. Warner suggested that the abundances of the radioactive isotopes K, U, and Th should also be similar on the two planets. The presence of granitic materials on the Venusian surface as indicated by one of the three determinations of K, U, and Th suggests that Venus and Earth had similar amounts of petrologic processing, and similar petrologic processing further suggests similar amounts of degassing. If indeed the two planets have had similar degassing histories, then the similar amounts of atmospheric ^{40}Ar on the two planets is confirmation that they have similar amounts of ^{40}K and hence all the radioisotopes. From those inferred abundances of heat sources and the known surface temperature, Warner constructed a suite of geotherms for Venus, the different geotherms being due to different assumptions concerning the amount of petrologic re-working of the Venusian crust and mantle. The geotherms were tested against Schaber and Boyce's (1977) observation that portions of the Venusian crust display large impact basins with a size frequency distribution similar to that observed for the Moon. To achieve the 10^9 year stability that those regions of the Venusian lithosphere must have had to preserve the large impact basins, the lithosphere must be thick relative to the lithosphere of Earth. If the mantle of Venus was hydrous all the calculated geotherms would

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yield a lithosphere with a thickness of 20 to 30 km. Such a thin lithosphere would not provide the required long-term stability. However, if the mantle of Venus was anhydrous, the lithospheric thickness would be over 50 km and the lithosphere would provide the long-term stability.

Water in Mars

Based on the observation of Moore et al. (1977) that the surfaces of large, igneous-looking rocks in the Viking landing site are hard and do not scratch or chip when "poked" by the teeth on the end of the Viking arm, and the observation of Huck et al. (1977) that those rocks are reddish indicating the presence of hematite, Warner (1979a) and Sato (1978) argued that the red color might be a primary feature of the rocks. If that is the case, then the red color is a probe to the source region in the Martian mantle where the lavas formed. Warner and Sato use a thermodynamic analysis of carbon oxidation to show that the oxidation state of lavas may be related to the water content of the mantle from which the lavas form. They argue that a magma is saturated with C and that the oxidation of the magma as it rises to the surface will be controlled by the equilibria of C - CO - CO_2. This reaction is calculated as a function of pressure and the fugacity of O_2 above the iron-wustite buffer. They argue that if the magma contains water the water will react with the C producing CO_2 and H_2. The H_2 will diffuse out of the system and the equilibria will be forced to the right. Once all available C is used in reaction with water, the magma is no longer confined to the C - CO - CO_2 equilibria curve, and it will rise to higher oxidation states. Thus the Moon which is anhydrous has lavas with very low oxidation states (near the iron-wustite buffer); the Earth which has a hydrous mantle has lavas that are more oxidized but still black (near the quartz-fayalite-magnetite buffer); and Mars may have a mantle with more water relative to Earth which leads to lavas that are oxidized enough to be red (near the hematite-magnetite buffer).

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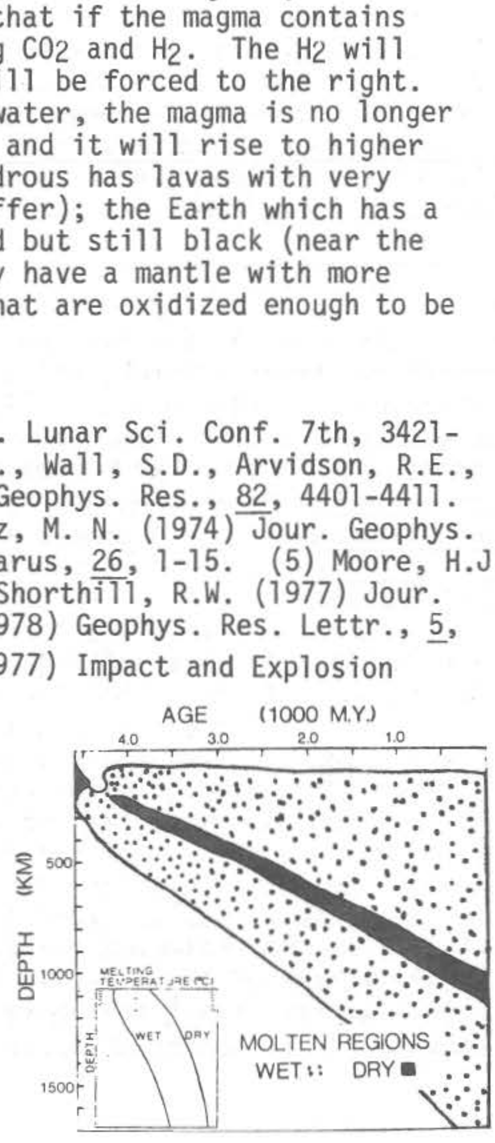

SIGNIFICANCE OF MEGASHEAR ZONES AND INTRACRATONIC BASINS IN THE EVOLUTION OF CONTINENTAL CRUST.

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A number of megashear zones in the North American continent have been postulated in recent years (1 - 5). They all share the following general characteristics: (a) Somewhere along their lengths they are defined by a linear trend in mapped faults, sedimentary basins, igneous intrusions, or known ore deposits. (b) They are for the most part, extrapolated across regions of unknown or uncertain geology. (c) Where defined on the basis of known geology, reactivation through geologic time can usually be documented. (d) Postulated models for their origin and the reasons given for their persistence through time are highly speculative. Although megashear zones such as those illustrated in Fig. 1 are being postulated in all continents, a basic question remains unanswered: Do they in fact exist or are they products of loosely constrained and over zealous interpretation of linear features on Landsat and other imagery? This question deserves serious consideration for the megashear zones postulated in exposed or near-surface Precambrian rocks because they may shed light on the tectonic evolution of continental crust in the Precambrian.

The existence of intracratonic basins in the North American continent is less speculative, but the understanding of their origin is as obscure as that of the megashear zones. Because such basins grow laterally with time (6), the age of the rocks and the nature of the unconformities along their margins are not a good guide to the geology of the original basins. It is suggested that a survey of available drill core and geophysical data is needed to advance the understanding of the processes which produced these significant additions to the Precambrian crust.

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ANORTHOSITIC COMPLEXES IN THE EARLY CRUST OF THE EARTH: COMPARISON OF MINERALOGY WITH LUNAR ANORTHOSITES; B. F. Windley, Dept. of Geology, University of Leicester, Leicester, England, F. C. Bishop, Dept. of Geological Sciences, Northwestern University, Evanston, Ill. 60201, J. V. Smith, I. M. Steele, R. C. Newton and J. S. Delaney, Dept. of Geophysical Sciences, University of Chicago, Chicago, Ill. 60637, and G. R. McCormick, Dept. of Geology, University of Iowa, Iowa City, Iowa 52242.

Anorthositic complexes in the early crust of the Earth. Anorthosite forms major stratigraphic units in extensive layered igneous complexes which occur in Archaean rocks of most continents. We consider here three classic examples: Fiskenaesset (W. Greenland), Sittampundi (S. India) and Limpopo (southern Africa). These complexes, up to 1km thick, were intruded into shallow-water sediments (now quartzite, marble and mica schist) and basic volcanic rocks (locally pillow-bearing) at -3.0±0.2Ga. They were subjected to variable high strain during thrusting, folding and refolding, and to pervasive recrystallization during regional metamorphism at amphibolite to granulite grade at ~10kb pressure in intermediate to lower levels of the Archaean crust. They now occur as conformable layers many tens to hundreds of kilometers long in granulite-gneiss belts. Where they were intensively intruded by tonalites, they occur as meter-size lenses and layers in deformed tonalitic gneisses.

Stratigraphically the complexes (1-4) tend to pass upwards from minor ultrabasic rocks and gabbros to prominent leucogabbros and anorthosites interlayered with chromitites. Modal variations of olivine, pyroxene, hornblende and Ca-plagioclase cover most rock types, but garnet is an abundant indicator of high-pressure metamorphism in the Sittampundi Complex. Oxides and sulfides are common accessories, especially in ultrabasic rocks.

From the mineralogical-petrological viewpoint, the important questions involve the effect of metamorphism and metasomatism on the original igneous differentiates, and from the geochemical viewpoint, the important questions involve the source of the magmas and what information is given on the crust-mantle development and its relation to planetary differentiation in general.

Fiskenaesset Complex. Reconnaissance of 63 highly metamorphosed rocks by light microscopy and electron microprobe and X-ray diffraction techniques (5-7) led to the conclusion that the silicate, oxide and sulfide minerals are variously-metamorphosed products in the granulite, amphibolite and (locally) greenschist facies of a moderately wet Al-rich basaltic magma which underwent crystal-liquid differentiation under fairly oxidizing conditions. The original igneous assemblage apparently involved early precipitation of Mg-Al-Ti-Cr-magnetite (hence the redox state), prolonged crystallization of highly calcic plagioclase and tschermakitic-magnesio-hornblende, and late crystallization of high-Fe, medium-Cr spinel.

Coexisting spinels outline a solvus between $(Fe,Mg)Fe_2O_4$ and $(Fe,Mg)Al_2O_4$, and wide tie-lines for Cr-poor specimens indicate equilibration below 500°C. Coexisting ilmenite and ferromagnesian silicates apparently equilibrated in most rocks near 650±100°C, but in the presence of late serpentine or tremolite, equilibration was probably below ~425°C. Coexisting pyroxenes apparently equilibrated to ~700-850°C depending on the laboratory calibration. Sulfide minerals have equilibrated as low as 230°C. Metamorphism definitely varies from place to place, and depends on the ease of equilibration for the various minerals.

The extent of metasomatism is unclear. Scattered biotite probably results from introduction of K, and amphibole rims around some pyroxenes require local introduction of H_2O. Nevertheless most features of the complex can be interpreted plausibly as the result of nearly isochemical metamorphism of an igneous complex with systematic cryptic variation of mineral composition. Huge primocrysts of Ca-rich plagioclase survive in rare rocks (8), and primary chemical zoning apparently remains: recrystallized regions have similar compositions to uncrystallized ones (9). The origin of clouds of amphibole inclusions in plagioclase megacrysts is uncertain – trapped magma, or exsolution from plagioclase?

Assuming isochemical metamorphism for most specimens, it is possible to attempt to reconstruct the original igneous mineralogy. The oxide assemblages are mostly the result of transformation of single phase Mg-Al-Ti-Cr-magnetite (lower zones) and Fe,Cr-rich spinel (upper zones). Using experimental data for ilmenite-silicate as a guide it was concluded that the present metamorphic "chromites" are less magnesian than the igneous precursors because of exchange with silicates. The Na-content of plagioclase tends to be higher where the amphibole/plagioclase ratio is higher, and this suggests that (i) much of the amphibole was originally igneous, and (ii) during metamorphism, Na is transferred locally from amphibole to plagioclase (incomplete study jointly with J. S. Myers: further work needed). If this is correct, the original plagioclase was mostly An_{85}-An_{95}.

Sulphides disseminated in the layered series are limited to the Ni-Co-Cu-Fe-S system and are concentrated in ultramafic zones and inter-layered lenses. The following were identified: pentlandite, pyrrhotite, chalcopyrite, pyrite, millerite, heazlewoodite, Co-pentlandite, polydymite, godlevskite, violarite, cubanite and digenite. The sulphides in the gabbros and anorthosites are Cu-rich with respect to those in the ultramafic rocks. Most of the S is probably magmatic, but some violarite and digenite may result from late alteration.

In contrast to the layered basaltic complexes which crystallized under reduced conditions (e.g. Skaergaard), the Fiskenaesset complex undergoes little enrichment to Fe-rich compositions, presumably because of early crystallization of magnetite under oxidizing conditions. Cr-rich spinel (chromite s.l.) crystallizes late, probably because amphibole rather than pyroxene is the dominant ferromagnesian mineral. Furthermore the plagioclase undergoes only moderate Na enrichment, presumably because of crystallization of Na-rich hornblende under wet conditions. All these factors fit together to indicate crystal-liquid differentiation from a wet, oxidized, high-Al basaltic liquid.

Limpopo Complex. Two sections of the complex reveal plagioclase, amphibole and spinel compositions which resemble those in part of the Fiskenaesset Complex. Extensive normal zoning of some plagioclases may result from Na-metasomatism, and alteration of amphibole and plagioclase to chlorite and epidote probably is associated with significant migration of mobile elements.

Sittampundi Complex. Extensive new data have confirmed the pioneering study of Subramaniam (10). Chemical compositions of chromite in chromitites resemble those of the Fiskenaesset Complex. Metagabbros record a granulite event at ~800°C and 12kb followed by retrogression which produced complex symplectites at ~650°C and 7-8kb. Aluminous rocks containing sapphirine and kyanite also indicate a complex history beginning with high pressure and temperature of metamorphism of adjacent Al-rich sediments.

General remarks. The mineralogy of the Fiskenaesset Complex resembles that of gabbros from the Peninsular Ranges batholith, California (11), and of peridotites (12) and gabbros from Connemara, Ireland (joint study underway with B. E. Leake). Apparently, the emplacement of igneous complexes with overall basaltic composition into granitoid terranes has taken place in

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continental margins throughout geological time (13). If so, the formation of these complexes is a function of tectonic position, and not of the time of emplacement. Since the Fiskenaesset complex dates some 1.5Ga after formation of the Earth, and some 0.8Ga after emplacement of the earliest surviving crust, analogies should not be drawn casually with lunar rocks believed to have formed from the primary differentiation of the Moon. Much detailed petrographic and geochemical work is needed to disentangle the effects of metamorphism and metasomatism before the properties of the terrestrial complexes are interpreted from a planetary viewpoint. It would be wise to estimate the bulk composition of a complex rather than relying on random samples. Finally it must be emphasized that anorthosite layers amount to only a small volume of the complexes, and that the complexes themselves are swamped by the granitoid rocks (*sensu lato*; mainly tonalitic gneisses) which surround them.

Comparison with lunar anorthosites. Unfortunately there is no stratigraphic control over lunar non-mare rocks, and a major effect is needed to extend the encouraging efforts to separate out the effects of meteoritic contamination (14) and to interpret the textures and mineral chemistry in terms of crystal-liquid differentiation and both shock and thermal metamorphism (14,15). Whatever will ensue, it seems quite safe to adopt the conclusion already reached from the Apollo 11 conference that an early differentiation of the Moon produced a complex crust composed of plagioclase-rich rocks and basalts. Such a differentiation involved anhydrous, reduced magmas, and must have been quite different from terrestrial differentiation which involved wet, oxidized magmas. As emphasized earlier, the Archaean anorthositic complexes (perhaps better called gabbroic complexes) do not result from an early Earth-wide differentiation, and cannot by any stretch of the imagination be explained by flotation of plagioclase in a magma ocean. Nevertheless the Archaean gabbroic/anorthositic complexes are worthy of detailed comparison with younger terrestrial complexes to determine whether the sources of the magmas (presumably the upper mantle) have changed with time.

The highly calcic nature of lunar plagioclase must be attributed to the low concentration of volatile elements in the bulk Moon, but the highly calcic nature of plagioclase in Archaean anorthositic complexes is probably the result of sequestration of most of the Na in igneous hornblende, whose crystallization depends on the availability of water. Certainly the calcic nature of the plagioclase does not reflect the composition of the bulk Earth since the dominant rocks in the crust were and are granitoids (s.l.) with feldspars of intermediate (Na+K)/Ca ratio.

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COMPILATION OF QUESTIONNAIRE

The questionnaire on the following page was distributed to the participants of the workshop. They were requested to complete it and return it to the conveners on the last day of the meetings. No attempt has been made to prioritize the responses; they are simply reproduced here from the questionnaires that were returned.

<u>QUESTIONNAIRE/WORKSHOP HOMEWORK</u>
(Due after discussion session) ____________________ Name (Optional)

1. What, in your view, are the (few) most fundamental questions and problems discussed?

2. What are the key observable parameters which are needed to address them?

3. What methods or approaches are likely to yield important results?

4. What <u>new</u> methods and approaches are likely to be important (including: a) coordinated work on sample suites, b(continental drilling, c) measurements from space — remote sensing, telemetered data, etc.)?

5. Specific recommendations for future research <u>programs</u> or <u>activities</u>.

6. Has workshop been profitable? How could it have been improved?

QUESTIONNAIRE/WORKSHOP HOMEWORK

1. What, in your view, are the (few) most fundamental questions & problems discussed?

   A. Relationship of crust-mantle development of the moon (anhydrous) to the earth (hydrous).

   B. The formation and nature of the primitive crusts.

   C. Generation of the lunar anorthosite, earth's early crust, impact processes vs. igneous processes.

   D. 1) Impact history of the moon and implication for early history of earth.
      2) Geochemical relationships between lunar and terrestrial rock.

   E. 1) What is the nature of protolith to early "grey" gneisses of Archean.
      2) Is concept of magma ocean viable for any or all terrestrial planets.
      3) To what extent are planetary histories equivalent, but more importantly are there also basic differences in their evolution history. (e.g. effect of H_2O on history)
      4) What is the effect of impact on early history of earth - nothing/everything - were they important for earth's evolution?
      5) What is the role of "plate tectonics" in early earth.

   F. 1) Nature of pre-four b.y. old terrestrial crust-ultrabasic, basic, intermediate, acidic, anorthositic, other?
      2) Structure and composition of crust of Venus.
      3) Origin of lunar highlands.
      4) Effect of four b.y. bombardment of earth:
         a. Initiate ocean basins.
         b. Initiate continental nuclei.
      5) Isotopic recycling a la Armstrong.

   G. Consideration of the origin and evolution of the earth's crust from a planetary point of view. Main question — at what stage of the planet's evolution did the continents as we know them originate?

   H. 1) How was crust put together? e.g. Role of vertical vs. horizontal tectonics, contribution of igneous additions to crust, source of igneous materials etc.
      2) If granulitic terrains represent deep levels of crust, what underlies them?
      3) Are Archean rocks proper analogies of the deeper crustal section?

   I. Are events recorded in early crustal development of other planets matched by any presently recorded events on earth?

   J. The question whether the oldest rocks exposed in the shields have been deformed under conditions of (*i*) crustal shortening, (*ii*) crustal extention, (*iii*) no bulk lateral strain of the crust.

Question 1 continued

K. 1) Obviously the return of the earliest terrestrial Archean crust.
   2) The role played by impacting in the early Archean crust.
   3) Implication of lunar magma ocean to other terrestrial planetary crusts.
   4) Thermal history of the crust.
   5) Crustal thickness and growth.
   6) What is below the 3.600 Ma crust now exposed at the surface?

L. Possible application of impact studies to early history of the earth.

M. 1) Presence, nature, and recognition of pre-greenstone belt crust.
   2) Bimodal nature of Archean magmatism, and its origin.
   3) Tectonic regimes in the Archean.
   4) Effects of initial impact on the earth.
   5) Stratigraphy of Archean sequences.
   6) Nature and origin of Archean lower crust.

N. 1) Formation of the continental crust of Earth.
   2) Formation of the crusts of other planets.
   3) Developmental processes of the Archean.
   4) Volatile inventory in the mantles of planets.

O. Hypotheses re <u>stages of development</u> of Archean crust.
   Hypotheses re <u>origin of terrestrial</u> and planetary crusts.
   Hypotheses re <u>stages of development</u> of planetary crusts.

* * * * * * *

2. What are the key observable parameters which are needed to address them?

   A. Contrasting crustal parageneses.

   B. Age dates, stratigraphy and petrology of crusts.

   C. Field geology, geochemistry, geophysics.

   D. Trace element geochemistry of oldest terrestrial rocks.

   E. 1) Physical properties of the mantle to plan constraints on correction and melting.
      2) Geochemical data on rocks derived by melting of mantle and crust to place limits on their history.

   F. Structural relations at depth between crustal components.

   G. Nature of earliest crust, chemical parameters, spatial distribution.

   H. We need some deep drill holes in the tonalitic gneiss domains, to test geophysical models. Also, we need mosfern (?) structural approaches in shield areas.

   I. Regions of early Archean crust in Labrador/Greenland, China, USA, Antarctica, South Africa, Zimbabwe, Australia, USSR (pre 3.5)

   J. More and better field data followed by integrat d geochemical and tectonic analysis. Integrated geophysical and geological investigations.

Question 2 continued.

K. 1) Precambrian geology (structural, peterogenesis,. . .) and geophysics (seismic profiling + force-field methods which can be airborne or other remote scanning);
   2) Planetary geophysical scanning + photogeology and remote sensing of chemical composition.

* * * * * * *

3. What methods or approaches are likely to yield important results?

   A. Experimental igneous petrology on simple silicate systems at high pressures.

   B. New sampling of course! Study of untouched lunar samples. Remote sensing of Mercury & Vanus. Isotopic studies of Archean igneous rocks.

   C. Integration of field geology, geochemistry, geophysics with models.

   D. 1) Creation of a reference Archean suite (?) (Chemical and pet.)
      2) Creation of well constrained thermal and thus mechanical models of early evolutionary stage of earth.

   E. 1) Seismic reflection studies of particular structures.
      2) Selected deep drilling.
      3) Integration of geophysical, geochemical, and field mapping results.

   F. Integrated studies of selected "type" areas and extent to which types can be extrapolated into other areas. This as opposed to setting up of model based on present clay tectonics.

   G. People working on the tectonics of the moon and other planets have to be content with interpreting the gross shape of major structures. This approach is hampered by the fact that any deformed body can reach its final geometry along an infinite number of kinematic paths.

   H. 1) Baseline geol. data in areas of proven early crust.
      2) Interlab comparison of age data on well collected material.
      3) Comparative field studies in all areas of pre 3.5 AE crust.
      4) Dissemination of information relating to other planetary crusts.

   I. Field investigations followed by integrated geochemical (particularly trace elements and isotopes), tectonic, stratigraphic, and geophysical investigations. Particular emphasis on the problem of alteration in interpreting geochemical data. (i.e. better understanding of Archean geology and processes)

   J. Synthesis of precambrian geology and planetary geophysical scanning + photogeology and remote sensing of chemical composition.

* * * * * * *

4. What new methods and approaches are likely to be important [including: a) coordinated work on sample suites, b) continental drilling, c) measurements from space — remote sensing, telemetered data, etc.]

   A. It is absolutely essential that a sound understanding of the thermodynamics of solid-liquid equilibria be developed.

   B. 1) Drilling of Archean basement.
      2) Detailed x-ray fluoresence survey of Mars, Mercury, & Venus.
      3) SEM petrography of areas of potential trapped liquid in primitive lunar anorthosites.

   C. In the Archean — deep geophysics, new geochem techniques (e.g., SM-ND).

   D. 1) Chem-petrological study of grey gneisses to attempt to determine protolith. A deep hole in such an outcrop area would also be of great help in terms of sampling.
      2) Co-ordinated (thermal models, mechanical models, fractionation models) into what effects of impact will be on 4.0 b.y. earth.

   E. 1) Lunar Polar Orbitor, stripped down version with only a) gamma-ray spectrometer, b) XRF, 3) TV imagery.
      2) Venus radar — VOIR.
      3) Continental drilling of diapiric quantities in Archean terrains: W. Australia, Minnesota, S. W. Canada.
      4) Sample return from Martian highlands.

   F. (a & b) Extensive seismic reflection work could provide a most useful framework in which to relate structural and compositional results.

   G. (b & c)

   H. 1) Deep drilling into early Archean crust of Greenland or Labrador.
      2) Coordinate work on samples from proven areas of the early Archean terrestrial crust.
      3) Experimental work on synthetic systems relating to granulite melting and generation of sialic magma in volatile rich environments esp. CO_2, F, Cl, etc.
      4) Use of multi-isotopic methods (comparatively) SM/Nd, Pb-Pb, Rb-Sr, K-AR.

   I. 1) Better analytical detail on properly collected sample suites. Amount of good field data outweighs analytical capabilities.
      2) Deep drilling would be important if it could penetrate close to Conrad, i.e. base of upper curst, but would require several holes in diverse geologic provinces e.g. granite batholiths, greenstone belts, gneiss belts.

   J. 1) Extend seismic exploration of the lithosphere into the region between the Moho and the mantle low-velocity zone.
      2) Extension of number of methods of penetrative remote sensing of earth and planets.

5. Specific recommendations for future research programs or activities.

A. 1) Isotopic and geochemical search for parent liquids of primitive lunar crustal rocks.
   2) Studies of continental drilling and xenolith samples.
   3) Continued study of terrestrial layered intrusions as analogues for lunar cumulate rocks.

B. Expand the geochem, dating, etc. to include as many elements as possible. Continue and expand field studies in the Archean Terranes.

C. It would be interesting to attempt some sort of quantitative treatment in an analysis of impact history. How much material was added, where is this material in the earth now, did it contribute to inhomogeneity in the mantle?

D. 1) Co-operative tests involving planetary and terrestrial geoscientists on the same team.
   2) Concurrent effort to define comparative thickness etc. of earth proto-crust and thus define initial conditions for continental (Archean) origin (e.g. Do impacts form continents or oceans?).

E. Detailed mapping of Dorsale Reguibat in North Africa.(?) Supurb (?) exposure of diapiric granites poorly mapped. Should be coordinated with Algeria, Mauritania, France.

F. More reflective profiling (of course).
   Xenolith studies.
   Most of all, interdisciplinary communication.
   More extensive use of geophysical information by geochemists, etc. and vice versa.

G. 1) Joint research on terrestrial Archean problems with interior (?) collaboration to resolve problems relating to the identification of possibly reworked early Archean crust, and apparent anomolies in some of the age methods.
   2) Field conferences in association with IACP?

H. 1) Continue study of earth anorthosites.
   2) Use facilities to curate meteorite collections, classical Archean locality specimens, etc.

I. U.S. COCORP and Canadian Co-operative Seismic Crustal Studies Group expand their activities (largely separate as to countries but some co-ordination is desirable.) Both are "under" Geodynamics Project. Action should take place under these international organizations.

6. Has workshop been profitable? How could it have been improved?

   A. Extremely profitable.
      It could be improved by a closer integration of the diverse topics.

   B. Yes. Better organization of sessions, i.e. Begin with overview talks and then go into more detailed talks; overview speakers should have had some communication with speakers to follow.

   C. Yes. Find money to invite foreign workers.

   D. Yes. Plate tectonics not well presented. An analysis from present back in time might be helpful.

   E. Very profitable. Concept of earth as a "planet" was well presented to Archean geologists. Could have gone on another day for more discussion and interaction. Major/trace element geochemists who work on proto-crust-Archean, should have been present.

   F. Extremely profitable and helpful.
      Improvements — Donuts not very good.

   G. Yes, it was profitable.
      Needed more opportunity for personal communication with individuals.

   H. Yes.

   I. Very worthwhile but I would have preferred more opportunity for general discussion as they came out.

   J. Yes.

   K. Certainly, because it brought together a diverse group of people.

   L. Yes.

   M. Yes. But could have had more discussion on some problems — particularly if discussion on groups of related papers could have been concentrated into a major discussion peirod. i.e. authors with different viewpoints could argue and refute each other's points.

   N. Yes, but a closer co-ordination of topics might have been desirable to make a more even emphasis; to allow more discussion; and to avoid overrunning of time and rushing of the last items. Was a good idea. Should be developed further.

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